THE JOURNAL OF GEOLOGY

A SEMI-QUARTERLY

VOLUME LIII

JANUARY–NOVEMBER, 1945

THE UNIVERSITY OF CHICAGO PRESS
CHICAGO, ILLINOIS

THE CAMBRIDGE UNIVERSITY PRESS, LONDON

PUBLISHED JANUARY, MARCH, MAY, JULY, SEPTEMBER,
AND NOVEMBER, 1945

COMPOSED AND PRINTED BY THE UNIVERSITY OF CHICAGO
PRESS, CHICAGO, ILLINOIS, U.S.A.

CONTENTS OF VOLUME LIII

NUMBER 1

NUMBER 2

NUMBER 3

NUMBER 4

NUMBER 5

NUMBER 6

VOLUME LIII NUMBER 1

THE JOURNAL OF GEOLOGY

January 1945

OUTLINE OF THE GEOLOGY OF CUBA

R. H. PALMER
Neptuno 1060, Habana, Cuba

ABSTRACT

Cuba is approximately 700 miles long and 50 miles wide. It is divided into eight physiographic provinces which roughly correspond to geological features. The geological column ranges from Middle Jurassic to Pleistocene. Much of the Middle and Lower Cretaceous is missing or has not been identified. There was extensive Cretaceous volcanism. The Upper Cretaceous and Tertiary have an abundant, well-preserved fauna. The Eocene and Oligocene faunas are Old World in contrast to the Miocene fauna, which is New World in relationships. The Tertiary is well represented from the Paleocene through the Miocene. The Upper Cretaceous and the Paleocene are land-derived sediments of unknown geographical origin. The remaining are largely marls and limestones.

There are both extrusive and intrusive rocks. Basic and acid rocks occur in both types.

There are various structures of diverse ages. In the west is a large overthrust followed by two large anticlines to the east. A large overthrust area including the Trinidad Mountains and intrusions occupy the central part of the island. A broad syncline with extensive intrusions forms the eastern end, with a mass of metamorphics lying at the extreme east.

Cuba in its present form dates from late Miocene. Previously the area was occupied by scattered islands or was completely submerged. During late Cretaceous and the Paleocene, portions of what is now Cuba were part of a large land mass that extended as far south as Jamaica.

Since assuming its present form, Cuba was submerged in the Pleistocene or Recent. This accounts for the paucity of land fauna on the island.

INTRODUCTION

The island of Cuba is 1,200 kilometers (720 miles) long with an average width of about 85 kilometers (50 miles). The eastern two-thirds (750 kilometers) lies in a straight line with a bearing of N. 70° W.; the following 200 kilometers lies in an east-west line, and the western 200 kilometers has a bearing of S. 65° W. The 80° meridian and the 22° parallel intersect near the center of the island. Cuba is bounded on the north by the Bahama Channel and the Gulf of Mexico and on the south by the Yucatán Basin and by the east end of the Bartlett Deep.

Recent submergence has left much of Cuba bordered by shallow water and numerous cays. Reference to the map (Fig. 1) shows that with an emergence of 8 fathoms these shoals and cays would become part of the mainland. The resulting outline would give a true picture of Cuba as an island mass surrounded by deep water.

TOPOGRAPHY AND PHYSIOGRAPHY

Cuba may be divided into eight physiographic provinces as follows:

1. The Organos Mountains that lie in the northern half of Pinar del Río Province. This range is 140 kilometers (85 miles) long and from 6 to 14 kilometers (3.5 to 8.5 miles) wide, with elevations to 1,500 feet. It consists essentially of a

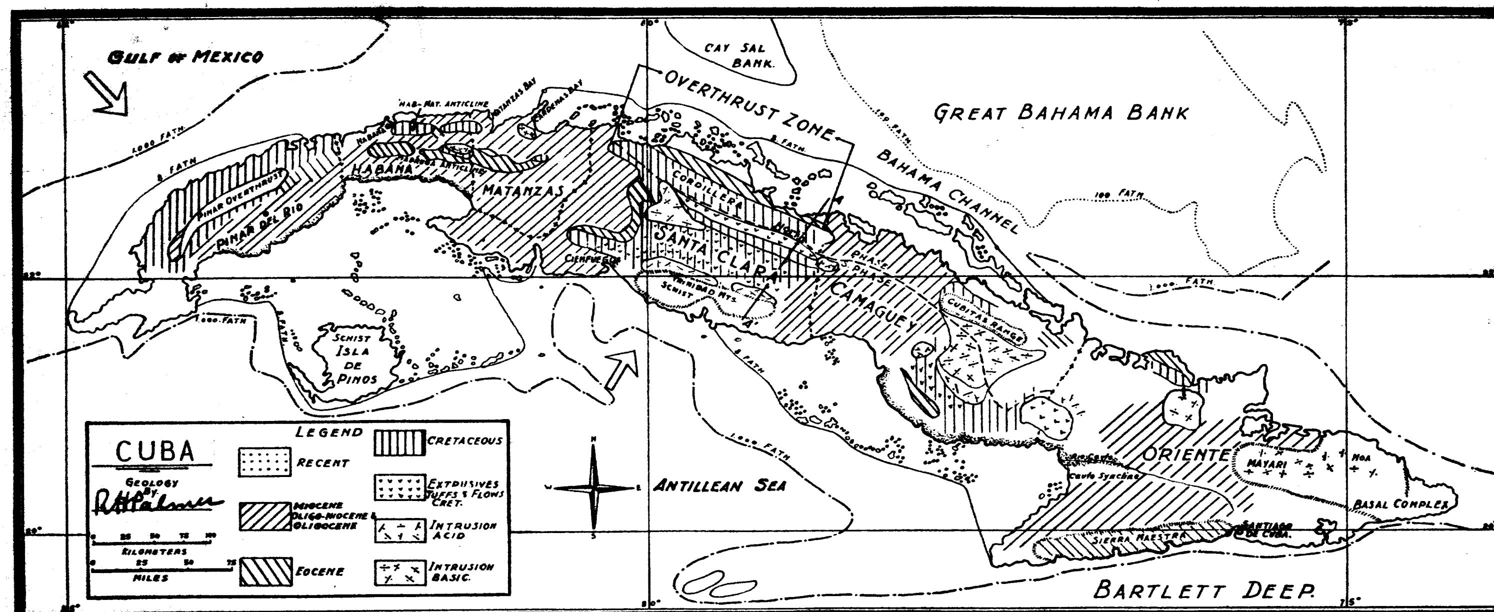


FIG. 1.—Map showing principal geological features and general distribution of Cretaceous, Tertiary, and igneous rocks in Cuba. Note re-entrant of 1,000-fathom line between Gulf of Mexico and the Pinar overthrust and a second between the Antillean Sea (Yucatán basin) and the Trinidad overthrust. The arrows indicate the direction of thrust (*Vergens* as used by Stille).

sheet of hard Cretaceous limestone overlying younger Cretaceous shales and sandstones of the Cayetano formation in an overthrust position. On the south side of the western end the range is broken up into large isolated blocks known as "mogotes." These mogotes are the most striking feature of the landscape of Viñales Valley (Fig. 2). Elsewhere, particularly in the eastern end of the range, erosion has left much larger masses of the overthrust sheet that completely cover the Cayetano except where the latter is exposed in the deeper stream beds.

The limestone mogotes support an abundant flora in spite of the scarcity of soil. The shales, on the contrary, have scant vegetation, but this is notable for pines and oaks which are rare elsewhere in Cuba and in striking contrast to the palms which elsewhere characterize Cuban scenery.

This province also includes the relatively high and deeply incised Cayetano shales area bordering the Organos Mountains on the south in a belt 8–12 kilometers wide.

2. The Cayetano Plain lies between the Organos Mountains and the north coast, west of the meridian passing through Pinar del Río City. It is a relatively small area 60 kilometers long by 10 kilometers wide and lies on the Cayetano formation. It is characterized by steep drainage courses near the mountains and by low, flat land nearer the coast. As on the outcrops of the same formation within the mountains, the vegetation is scanty.

3. This is the folded zone occupying the northern half of Habana Province and extending eastward to about the middle of Matanzas Province. It includes the Habana-Matanzas and Madruga anticlines and the Almendares–San Juan syncline between them. Along both anticlines erosion has removed the Tertiary limestone and exposed the softer Cretaceous shales, producing a topography of low, rolling hills flanked on either side by Tertiary limestone cliffs. The soil derived from the Cretaceous is considered mediocre in Cuba. Cane is its principal crop.

4. South of the Organos Mountains and the folded zone in Habana and Matanzas provinces and extending eastward to about the Cienfuegos meridian in Santa Clara Province is the flat Coastal Plain. In western Matanzas Province an arm of this plain crosses the island to the north coast. This physiographic province is a monotonous plain characterized by red soil, many sinkholes, and underground drainage. A belt of sand and terrestrial debris borders the south coast. The red soil is derived from the underlying Güines limestone and, even where there is only a shallow accumulation, produces the finest cane land in Cuba. Where the rock is at or near the surface, precipitation quickly enters the porous limestone resulting in desert conditions and a xerophytic flora.

5. Santa Clara (or Las Villas) Province[1] forms a single, complicated physiographic province. A zone of overturned folds and overthrusts, called the "Cordillera," lies along the north coast. This has been reduced to low ridges by erosion. South of the Cordillera is a large area characterized by several types of intruded rocks that stand out as hills up to 600 feet in height. In the southern part of the province are the Trinidad Mountains, with peaks nearly 3,000 feet high.

6. A second coastal plain occupies the

[1] The official name of Santa Clara Province has been recently changed to Las Villas. Since the current maps all bear "Santa Clara," that name has been used here.

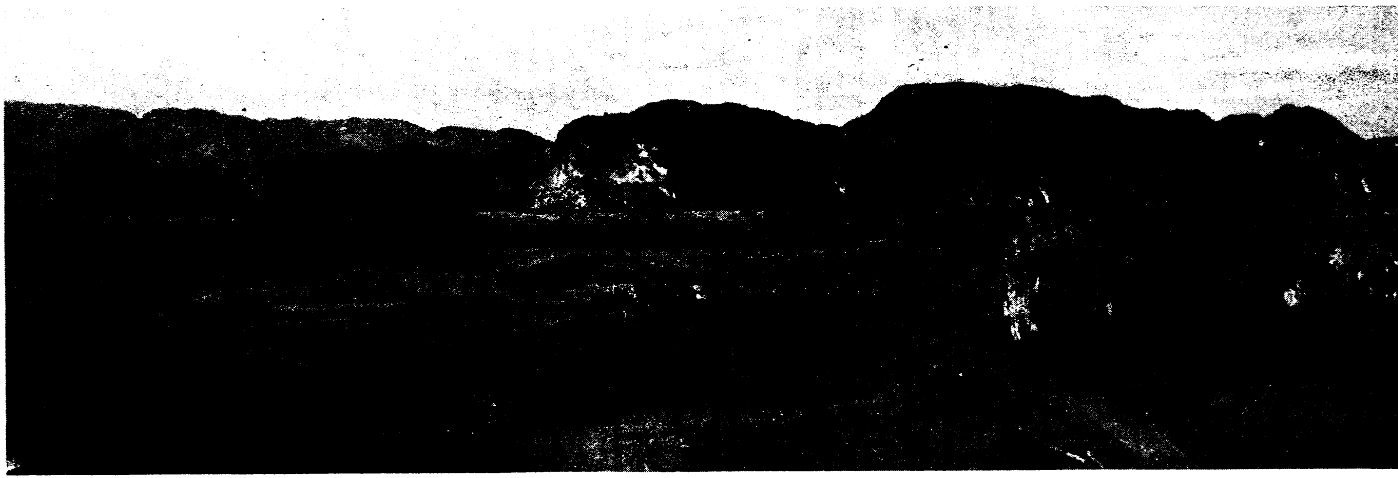


FIG. 2.—Looking north across Viñales Valley. The "mogotes" (isolated hills) are remnants of the dissected Viñales limestone sheet thrust from the northwest over the Cayetano shales that form the valley floor. (Photograph by Dr. Roberto Machado.)

southern half of Camagüey Province and extends eastward across the Cauto Valley to the Sierra Maestra in Oriente Province. As in the western Coastal Plain, a broad arm crosses the island to the north coast. This lies between the Cordillera and the meridian passing through Camagüey City. It also resembles the western Coastal Plain in being underlain for the most part by the same Güines limestone capped by the same rich soil. So similar are the two coastal plains that they may be considered a single unit interrupted by the Trinidad Mountains.

7. The folded and intruded area in northeastern Camagüey and Oriente provinces.

8. The Sierra Maestra in the southern part of Oriente Province and the mountainous zone in the eastern portion of the same province.

TABLE 1

COLUMN*

	Series	Formation		Description	
Tertiary	Miocene	Matanzas		marl	
				unconformity	
		La Cruz		marl	
	Oligo-Mio.	Güines		limestone	
	Oligocene	Cojimar		chalk	
				unconformity	
		Tinguaro		marl	
	Eocene	Príncipe		chalk	
		Universidad	Tuff Series	chalk	
				unconformity	
	Paleocene	Capdevila		sh, ss, cgl	
Mesozoic	Cretaceous	Habana		ss, sh, ls, cgl, tuff	Land derived
		Cayetano		sh, ss	
		Provincial		ls with sh beds	
				unconformity	
		Aptychus Beds (Viñales)		ls, few shales	
	Jurassic	Quemado		ls	
		Jagua		schistose ls	
	?	Santa Fé Schists, Gerona Marble, Trinidad Schists			
	?	Basal Complex in Oriente (Taber)			

*The principal formations only are given in the column. Unimportant formations, faunal zones, and other small subdivisions are omitted.

These eight physiographic provinces correspond for the most part to geological features that will be described later.

PALEOZOIC?

In the eastern end of Cuba there is an area of badly sheared and contorted

schists which S. Taber[2] calls the "Basal Complex" and believes to be Paleozoic. The age determination is based entirely on lithology, as no fossils have been found in the series.

SANTA FÉ SCHISTS, GERONA MARBLE, AND TRINIDAD SCHISTS

In the Trinidad Mountains of southern Santa Clara Province there is a thick series of hornblende, micaceous, and calcareous schists. On the Isle of Pines there is a similar series of hornblende, micaceous, and quartz schists and phyllites with a thick limestone member that is altered to marble. The schist series on the Isle of Pines, C. W. Hayes, T. W. Vaughan, and A. C. Spencer[3] named the "Santa Fé schists" and the limestone the "Gerona marble." Many lithologic similarities have suggested tentative correlation of the two metamorphic areas.[4] Neither series is fossiliferous, hence their age is not known. The opinion has also been expressed that the marbles and schists of the Isle of Pines are the metamorphic equivalents of the Viñales limestone and the Cayetano shales in the Organos Mountains of Pinar del Río Province,[5] but no supporting evidence has been offered. Rutten[6] bases the correlation on lithologic grounds. The Pinar del Río series has not been metamorphosed except locally where shearing has been intense or near intrusions, while on the Isle of Pines the entire mass has been altered to marbles and schists. However, the thinly bedded, 34,500 feet of Cayetano shales with thick lenses of sandstone are suggestive of the thinly bedded micaceous and quartz schists of comparable thickness[7] of the Isle of Pines. This is particularly true of the siliceous phase of the Cayetano nearest the Isle of Pines toward the southwest end of the Organos Mountains.

JURASSIC

JAGUA FORMATION

The oldest formation in Cuba whose age determination is based on fossil evidence belongs to the Jurassic. This is a very thinly bedded, shaly limestone of nearly schistose structure that outcrops at the base of some of the northern mogotes in Pinar del Río Province. It has a thickness of about 400 feet. Within the limestone are numerous concretions locally known as "jicoteas" (turtles) or "quesos" (cheese). These concretions contain fish remains that have been identified as Oxfordian and ammonites that range from Bajocian to Portlandian.[8] Within the chambers of the ammonites there is often liquid or dry asphalt.

In 1935 R. E. Dickerson and W. H. Butt[9] found ammonite-bearing concretions in the thinly bedded Jagua formation which they mistook for the Cayetano formation which it somewhat resembles. Subsequent work has shown that the Viñales limestone and occasionally portions of the underlying Jurassic Jagua formation have been thrust over the Cayetano, with the result that in

[2] "Sierra Maestra of Cuba," *Bull. Geol. Soc. Amer.*, Vol. XLV (1934), pp. 567–619, Pls. 57–85.

[3] *Informe sobre un reconocimiento geologico de Cuba* (Habana: Sec. Agric., Dirección de Montes y Minas, 1938).

[4] L. Rutten, "Geology of the Isle of Pines," *Proc. K. Akad. Wetensch. Amsterdam*, Vol. XXXVII (1934), pp. 3–8.

[5] Barnum Brown and M. O'Connell, "Correlation of the Jurassic Formations of Western Cuba," *Bull. Geol. Soc. Amer.*, Vol. XXXIII (1922), pp. 639–64; J. W. Lewis, "Geology of Cuba," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XVI (1922), pp. 533–55.

[6] P. 6 of ftn. 4 (1934).

[7] *Ibid.*, p. 5.

[8] Brown and O'Connell, ftn. 5 (1922).

[9] "Cuban Jurassic," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XIX (1935), pp. 116–18.

these places the Jagua appears to be directly above or part of the Cayetano (Fig. 3).

The formation is known only in the Organos Mountains in the western part of the island. The best-known exposure is in the area known as Jagua Vieja, 3 kilometers east of Constancia and 10 kilometers northeast of the village of Viñales. "Jagua" is, therefore, suggested as an appropriate name for this formation.

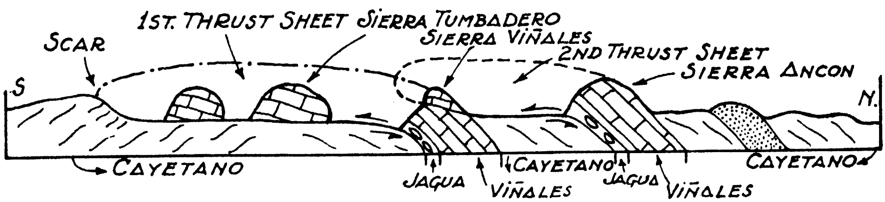


FIG. 3.—Northwest-southeast section through the Organos Mountains. The "scar" on the south marks the south edge of the overthrust.

QUEMADO FORMATION

The next younger formation occurs locally in Santa Clara Province. It is a series of hard, brown, siliceous and calcareous sandstones and hard, brown limestone. A scant, poorly preserved ammonite fauna occurring in this formation has heretofore been identified as Lower Cretaceous. In 1942, R. Imlay[10] described this fauna and determined it as Portlandian Jurassic.

The fauna described by Imlay from La Catalina, Pinar del Río Province, and Loma Camajan, Camagüey Province (loc. 18581), is believed to belong to this formation. Jaworski, as quoted by L. W. J. Vermunt,[11] believes the La Catalina fauna to be Tithonian or more probably Lower Cretaceous in age.

As this limestone has its greatest development near Quemado de Güines in Santa Clara Province, it is given the name "Quemado formation." Fourteen kilometers west of Sagua and directly south of Caguaguas this formation makes up the terrain for a distance of 5 kilometers (3 miles). Here it has a thickness of 4,400 feet.

CRETACEOUS

APTYCHUS BEDS—VIÑALES LIMESTONE

Lying directly above the Quemado formation in Santa Clara and Camagüey provinces are the Aptychus beds, so called for the abundant aptychi or ammonite opercula they contain. This formation is of long longitudinal distribution, occurring in the northern half of Cuba from Pinar del Río to Camagüey, a distance of 720 kilometers (430 miles). Good exposures occur in the eastern end of the Organos Mountains in western Cuba. In Santa Clara Province one of the best exposures is north of Loma Penton, 10 kilometers west of Sagua la Grande. Here 1,300 feet is exposed above the Quemado. The long row of hills 5 kilometers west of Camajuaní, Santa Clara Province, is of Aptychus beds, as is also the greater part of Loma Camajan in eastern Camagüey.

[10] "Late Jurassic Fossils from Cuba," *Bull. Geol. Soc. Amer.*, Vol. LIII (1942), pp. 1417–78.

[11] "Geology of the Province of Pinar del Rio, Cuba," *Geog. Geol. Mededeelingen, Phys.-Geol. Reeks No. 13* (1937).

Throughout its extent this formation is strikingly uniform in character. On the surface it is a fine-grained, brownish, thinly bedded limestone. The strata vary from paper thinness to several inches in thickness. In addition to the aptychi, it contains a few ammonite molds and occasional fish skeletons and scales and rare mollusks. Radiolaria abound in many exposures. Locally throughout its distribution there are thin beds of black chert that replace the limestone.

The Aptychus beds weather to a residual red clay with flat, platy boulders. Where not too greatly reduced by erosion, they form rounded hills. The red soil is comparable both in appearance and in fertility to the red soil derived from the Güines limestone. In fact, the red soil derived from both these formations has been mapped as Matanzas Clay by H. H. Bennett and R. V. Allison.[12] At depth the Aptychus beds are bituminous, shaly, soft marlstone. Their bituminous nature, together with the oil seeps occurring in them and in the overlying formations, has led to the conclusion that this formation contains the source beds of oil and asphalt in Cuba.

In many localities throughout its distribution there are thinly bedded, black, shaly cherts lying below the aptychus-bearing member. Below these and at the base of the formation and equally as spotted in occurrence is a coarse, calcareous sandstone or fine conglomerate. The best-known exposure of the three members of the formation is at Santa Fé near Camajuaní, on the road to Santa Clara in Santa Clara Province. At Cenral Zaza, near Placetas, Santa Clara Province, thin beds of tuff occur in Aptychus beds. M. G. Rutten[13] reports a similar observation.

The typical Aptychus beds do not extend westward beyond San Diego de los Baños in Pinar del Río Province. Here the Viñales limestone appears to occupy the same stratigraphic position. This is a hard, brittle, gray, massive limestone in which fossils have not been found. It forms the mass of the Organos Mountains and weathers in large, steep-sided blocks which are the above-mentioned mogotes (see p. 3) (Fig. 2). A review of the descriptions of the Viñales limestone and the Aptychus beds reveals that the qualities of the two are in distinct contrast. Scarcely a descriptive term of one can be applied to the other, though the two are probably in part equivalent. This as a practical measure warrants the use of two formational names.

The age of the Viñales and Aptychus beds has been a matter of much discussion. Hayes, Vaughan, and Spencer[14] considered the Viñales Paleozoic. E. DeGolyer[15] described the former and ascribed it to the Jurassic on the basis of the ammonites found associated with the mogotes. Later, Brown and O'Connell[16] also considered the Jurassic ammonites as coming from the limestone talus at the base of the Viñales limestone cliffs. J. W. Lewis[17] considered the Viñales to be Jurassic. Dickerson and Butt[18] found an ammonite fauna near

[12] *The Soils of Cuba* (Washington: Tropical Plant Research Foundation, 1928).

[13] "Geology of the Northern Part of the Province of Santa Clara, Cuba," *Geog. Geol. Mededeelingen Phys.-Geol. Reeks No. 11* (1936).

[14] P. 21 of ftn. 3 (1938).

[15] "The Geology of the Cuban Petroleum Deposits," *Bull. Amer. Assoc. Pet. Geol.*, Vol. II (1918), pp. 133–67.

[16] P. 648 of ftn. 5 (1922).

[17] "Geology of Cuba," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XVI (1932), pp. 533–55.

[18] P. 117 of ftn. 9 (1935).

La Catalina, Pinar del Río Province, in what they believed to be Viñales limestone. Subsequent field work has shown that these fossils come from beds probably equivalent to and herein named "Quemado formation" below the Viñales or Aptychus. The fossils are Portlandian in age, according to Imlay.[19] F. Trauth,[20] on the basis of the aptychi, ascribed the aptychus-bearing beds to the Lower Cretaceous Neocomian.[21] Imlay[22] uses the term "Viñales" to include both the typical Viñales limestone and the lithologically distinct Aptychus beds and ascribed it to the Jurassic on the assumption that all the ammonites he describes were from the Viñales formation. But, in fact, the faunas of three formations were described or mentioned in his paper: (1) the Jagua formation, Oxfordian (?), lowest Jurassic which contains ammonite-bearing concretions; (2) the Quemado formation, which furnished the diagnostic fossils of the Portlandian Jurassic; and (3) Aptychus beds which lie immediately above the Quemado formation. The age determination of the Aptychus beds, therefore, remains unchanged and is still open to question.

From the eastern half of the Organos Mountains eastward, the Aptychus beds are generally calcite-veined and much more contorted and sheared than the younger Cretaceous and Tertiary formations, indicating that they were subjected to marked diastrophism prior to the deposition of later sediments. This was the first and the greatest diastrophic event in the history of Cuba of which there is record. In the western 100 kilometers of the island the Viñales underwent the same compression, but it was thrust bodily over younger formations without being sheared except at the base.

[19] P. 1442 ff. of ftn. 10 (1942).

[20] "Ueber Aptychenfunde auf Cuba," *Proc. Akad. Wetensch. Amsterdam*, Vol. XXXIX (1936), pp. 66–76.

[21] Trauth (*ibid.*) reports that several species of Cuban aptychi were earlier described from the Cape Verde Islands, from the Alpine-Mediterranean region including Spain, southern France, the Alps, the Carpathian Mountains, and Algiers. He has therefore placed the Cuban species in synonomy with them. He mentions that the Cape Verde Island aptychi are from platy limestone (*Plattenkalken*). Dr. E. Jaworski of Bonn University has examined material from the Cape Verde Islands and from Cuba. He stated (verbal communication) that the lithology of the two is the same and that the aptychi are unquestionably Barremian (Lower Cretaceous).

Some years ago an opportunity was presented to examine material collected from Persia by the Anglo-Iranian Oil Company. The thin-bedded, marly limestone, tawny or buff in color with what appeared to be Radiolaria, was a duplicate of the Aptychus beds of Cuba. I was informed that the formation carried aptychi and was ascribed to the Neocomian. There is thus strong evidence that during the Neocomian the Mediterranean (Tethyan) Sea extended from Cuba eastward 5,000 miles to Rumania with the suggestion that it reached Persia, 2,000 miles still farther east.

[22] P. 1435 of ftn. 10 (1942).

PROVINCIAL LIMESTONES

The known Cretaceous record from the Aptychus beds to the Habana formation is very incomplete. A number of limestone lenses occur in a thick tuff series in southern Santa Clara and Camagüey provinces. Some of these are referred to the Provincial limestone; others are younger. A. A. Thiadens[23] described the Provincial limestones as follows: "The Provincial limestones are a light yellow, dark gray, grayish blue, fine to medium grained, mostly microconglomeratic rocks, cut by many small and large, white calcite veins. They are thick bedded but mostly they are thin bedded." The formation carries a rather rich though poorly preserved fauna of caprinids, Nerineidae and an occasional Apricardia, which indicate Cenomanian-Turonian age.

[23] "Geology of the Southern Part of the Province of Santa Clara, Cuba," *Geog. Geol. Mededeelingen, Phys.-Geol. Reeks No. 12* (1937).

CAYETANO FORMATION[24]

The Cayetano formation has not been recognized outside Pinar del Río Province. It reaches its greatest development north of the Organos Mountains and west of the meridian passing through Pinar del Río City. Here 34,500 feet[25] is exposed. As far as known, this is the thickest formation in Cuba. It extends around the west end of those mountains and eastward along their south flank, occurs under the mogotes within the Organos Mountains, and occupies the greater part of the northern half of the province. It is exposed as windows where the overthrust Viñales limestone is a more or less continuous sheet, as in parts of the eastern end of the range. The Cayetano is a thick series of very dark greenish sandstones and dark micaceous shales that weather to various tints and shades of red and brown and even white. The weathered-surface exposures bear little or no resemblance to the same beds below the oxidized zone. The Cayetano has been referred to in published reports as a series of schists and phyllites. Schuchert[26] speaks of the (San) Cayetano as metamorphosed strata that make up the basement complex. Lewis[27] calls it the "Pinar schist" and describes it as yellowish, friable, mica schists and reddish shale slates. These terms are quite misleading, as they imply metamorphism. Personal observation witnesses no metamorphism except near igneous intrusions or where the rocks were exposed to local stresses incident to the overthrusting. The only alteration shown by the great mass of Cayetano shales and sandstones is due to weathering.

The Cayetano is largely shale, though there are well-defined sandstone lenses, particularly on the north side of the Organos Mountains. Where the latter are of considerable magnitude they form prominent ridges up to several kilometers in length. Near the base of the formation and rarely higher in it there are a few hard, clastic, or massive limestone lenses. Usually the limestone has a bituminous odor, and occasionally oil seeps are seen.

The Cayetano soils are highly siliceous and, except in a few small areas where the limestone is present, are very poor and but little cultivated. As already stated, the formation supports a scant flora, a notable feature of which is pines and oaks. The oaks have not been noted in Pinar del Río except on this formation or its derivatives. Pines occur both on the Cayetano and on the iron-rich residual, lateritic soils of the serpentines. When supplied with sufficient moisture, the Cayetano soil produces the famed tobacco of Vuelto Abajo.

The age of the Cayetano formation is not satisfactorily known. It lies above the Viñales limestone (Aptychus beds) and below the Big Boulder bed of the Habana formation. DeGolyer,[28] who described it, called it "Cretaceous"; Brown and O'Connell[29] consider it pre-Oxfordian; Lewis [30] called it the "Pinar schists" and considered it pre–Middle

[24] The formation was described by DeGolyer (p. 140 of ftn. 15 [1918]) and given the name "Cayetano." Dickerson and Butt (pp. 116 ff. of ftn. 9 [1935]) referred to it erroneously, using the name "San Cayetano." This error was followed by C. Schuchert and by Imlay (pp. 1421 ff. of ftn. 10 [1942]). (*Historical Geology of the Antillean-Caribbean Region* [New York: John Wiley & Sons, 1935], pp. 410, 495, 515, etc.)

[25] I am indebted to E. N. Pennypacker for this calculation. It was taken from a section from the central part of the Organos Mountains northwest to the coast.

[26] P. 495 of ftn. 24 (1935).

[27] P. 534 of ftn. 17 (1932).

[28] P. 150 of ftn. 15 (1918).

[29] P. 648 of ftn. 5 (1922).

[30] P. 534 of ftn. 17 (1932).

Jurassic. Metcalf, in a discussion of Lewis' paper,[31] said he believed it to be above the Viñales. Dickerson and Butt[32] ascribed it to the Jurassic, believing that the Jurassic ammonite-bearing concretions were indigenous to it. H. M. E. Schürmann[33] places the phyllite (Cayetano) stratigraphically below the Jurassic. Vermunt[34] placed the Viñales and Cayetano together in one formation—the San Andres. This formation also includes the Quemado. Referring to the San Andres, he states: "We take the phyllitic, quartzitic rocks and the limestones to belong to a continual sedimentation cycle in Jurassic to lower Cretaceous time."[35] Imlay[36] quotes Dickerson to the effect that the high fixed-carbon ratio of the bitumen often found in the chambers of ammonites from the concretions presumed to have been taken from the Cayetano formation are an indication of age. This is misleading, as the ammonites in fact come from the Jurassic, and, furthermore, bitumen is merely dried petroleum which in the tightly sealed chambers is often found as a liquid; where evaporation has been possible the volatiles have left, giving a higher percentage of fixed carbon in the residue.

The only fossil remains in the typical Cayetano are small clams resembling the genus *Sphaerium*, which have been found in a few localities, and a few vegetal remains. In the limestone lenses, however, near the base of the formation there are a few Foraminifera. As these have not been studied in detail, nothing can be stated except that the genera appear to be the same as those occurring in the Lime gravel member of the Upper Cretaceous Habana formation. It is these limestone lenses in the Cayetano that Vermont calls the "Mountain facies" of the Habana formation.

[31] *Ibid.*, p. 553.

[32] P. 117 of ftn. 9 (1935).

[33] "Massengesteine aus Cuba," *Neues Jahrb. Beilage-Band 70*, Abt. A (1935), p. 336.

[34] P. 7 of ftn. 11 (1937).

[35] *Ibid.*, p. 11.

[36] P. 1422 of ftn. 10 (1942).

TUFF SERIES

A thick series of tuffs, agglomerates, and flows makes up a large portion of the Cretaceous column. These range from a few thin beds in the Aptychus to the top of the Cretaceous (see Table 1) and in Oriente Province to the Middle Eocene. The series has its greatest development in southern Santa Clara Province in central Cuba. Here it forms a belt 120 kilometers (70 miles) long by 25 kilometers (15 miles) wide and, according to Thiadens' estimate, has a thickness of 8,000 meters, though it now is believed to be much thicker. In Matanzas Province more than 2,000 feet is known; in Habana 5,000–6,000 feet; and north of the Organos Mountains in western Cuba, 1,300 feet is exposed. In Camagüey Province the thickness is comparable to that of Santa Clara. The tuffs have been encountered in oil wells in Habana and Matanzas provinces.

Except in central Cuba, the tuffs are not everywhere continuous. In Habana, for example, there is a tuff lense entirely surrounded by normal marine sediments, indicating that it is the product of a single volcano.

The ejected material consists of tuffs, agglomerates, and flows. These have been excellently described by Thiadens,[38] who named the formation. The lithology of the series is of the intermediary type, and attention may be called to the fact that none of the known examples is suf-

[37] P. 12 of ftn. 23 (1937).

[38] *Ibid.*, p. 12.

ficiently basic to be the parent-rock of serpentine. This suggests that the extensive serpentines of Cuba are genetically unrelated to Cretaceous igneous activity.

The widespread occurrence of Radiolaria and Foraminifera in the tuffs and agglomerates indicates that a large portion of them was deposited in the sea. Many lenses of marine limestone, varying from a few feet to several hundred, are indicative of the same origin. Several of these lenses are referred to the Provincial limestone described above.

HABANA FORMATION[39]

This widely distributed formation occurs in all the provinces of Cuba. It carries an abundant and well-preserved fauna of foraminifers, corals, mollusks, and echinoids that is Maestrictian, Upper Cretaceous, in age. The Habana formation is well exposed along the axes of both the Habana-Matanzas and the Madruga anticlines. In central Cuba several hundred square miles of Upper Cretaceous deposits are exposed on the Rodas anticline (see p. 25).

There are three members in this formation in the northern half of the three western provinces. The oldest is termed the "Lime gravels," composed of loosely consolidated limestone pebbles which vary in size from coarse sand to half an inch in diameter. The gravels contain orbitoids, rudist fragments, and, rarely, echinoids. Within the pebbles themselves there are often alveolinellid Foraminifera which are also considered Upper Cretaceous in age.

The second member is the "Cone sandstone," so called from the cone-shaped concretions that form under subaerial weathering. These cones develop in an inverted position with the base parallel to the bedding plane. This member is a light-gray, uniformly fine-grained, calcareous sandstone with scattered chlorite grains and lime cement. Both the Lime gravel and Cone sandstone are well exposed at San Francisco on the Central Highway, 12 kilometers southeast of Habana. Locally the Cone sandstone varies laterally to white marls. Its hardness makes it a valued stone for building and road purposes. Topographically it forms hills.

The third and youngest member of the Habana formation is the "Big Boulder bed," named from the habit of weathering to a residue of large boulders that makes its identification in the field possible even at a distance. The boulders are in part from conglomerates in the formation and in part from the disintegration of the limestone beds. The excellent Upper Cretaceous molluscan, coral, foraminiferal, and echinoid fauna with few exceptions is confined to this member. Very rarely large fragments of silicified wood in the form of rounded boulders are encountered. Though this member is well represented in both the large anticlines of western Cuba, the fauna is very largely confined to the southern one. On the other hand, tuffs are by far the more abundant in the northern structure, where fossils are few.

The three members occur together at Cantarana, 9 kilometers east of Madruga, Habana Province. They were, however, not recognized in oil wells located at a considerable distance south of their surface exposures in Habana and Matanzas provinces, but instead there was a dark-gray shale that is their equivalent. On the other hand, Big Boulder bed foraminifers and Cone sandstone

[39] The Upper Cretaceous and Tertiary of Habana and Matanzas provinces have been studied and ably described by Ing. Jorge Brodermann, "Determinación geológico de la cuenca de Vento," *Tercer Congreco Nacional de Ingeniería* (1940), pp. 8–28.

were encountered in a well near the north coast a short distance east of Habana. The data seem to indicate that the three members of the formation were not deposited as such except in the general area mentioned.

The two lower members of the Upper Cretaceous are confined to the three western provinces. The Big Boulder bed is the sole representative of the Upper Cretaceous occurring in all the provinces. The three members are shallow-water deposits with the exception of the white marls, which are believed to be the lateral facies of the Cone sandstone. In southern Santa Clara and Camagüey provinces, both the Middle and the Upper Cretaceous are tuffs with clastic limy lenses.

TWO PHASES OF UPPER CRETACEOUS

There are two phases of the Upper Cretaceous (see Fig. 1), as pointed out by M. G. Rutten.[40] The marls, sandstones, limestones, conglomerates, and igneous debris already mentioned as occurring in Pinar del Río, Habana, Matanzas, and the southern half of Santa Clara and Camagüey provinces constitute the southern phase as described by Rutten. The name "Habana formation" usually refers to this phase. The northern phase is limited to the northern half of Santa Clara, Camagüey, and Oriente provinces. This latter phase is almost a pure limestone.

JARONÚ LIMESTONE

The best-defined exposure of the northern phase is on the lands of the Jaronú sugar mill in Camagüey Province, where a section of 27,000 feet was measured. Accordingly, the name "Jaronú limestone" has been applied to this phase. The presence of Barrettia, caprinids, and Nerinea low in this section indicates that some of the northern phase is older than Upper Cretaceous.

[40] Pp. 21, 23 of ftn. 13 (1936).

LIMESTONE BRECCIA

Near the top of the northern phase in the vicinity of Camajuaní is a thick limestone conglomerate carrying large, sharply angular fragments of Aptychus limestone and chert. What appears to be the same conglomerate appears near the top of the Jaronú section. The angular boulders suggested the name "Limestone breccia" for this member of the Jaronú limestone. This member carries Foraminifera of the species occurring in the southern phase of the Upper Cretaceous.

Small chloritized fragments of tuff also occur in the Limestone breccia. These are the only evidence of Upper Cretaceous volcanic activity in the northern phase.

Although at least in part the same age, the northern and southern phases were deposited under entirely different conditions of sedimentation and evidently in widely separated areas. There is some 27,000 feet of limestone lying above the Aptychus beds on the north and a comparable amount of tuffs occupying the same stratigraphic position to the south, and yet the two are separated by a distance of only 5–10 kilometers (see Fig. 5). It appears that the southern representative was pushed northward during the extensive overthrusting that will be discussed later.

The Big Boulder bed presents several problems. The boulders from the conglomerate phase are of types whose parent-formation is nowhere known in Cuba. Where similar types of rocks do occur, they are several hundred kilometers distant and belong to periods of igneous activity long subsequent to the Creta-

FIG. 4.—Viñales Valley looking southwest. The cliff is the south side of Sierra de Viñales. The portion below the white line resembling talus is the remnant of an earlier overthrust sheet of Viñales limestone; that above is a later overthrust sheet. See section, Fig. 3. (Photograph by T. E. White)

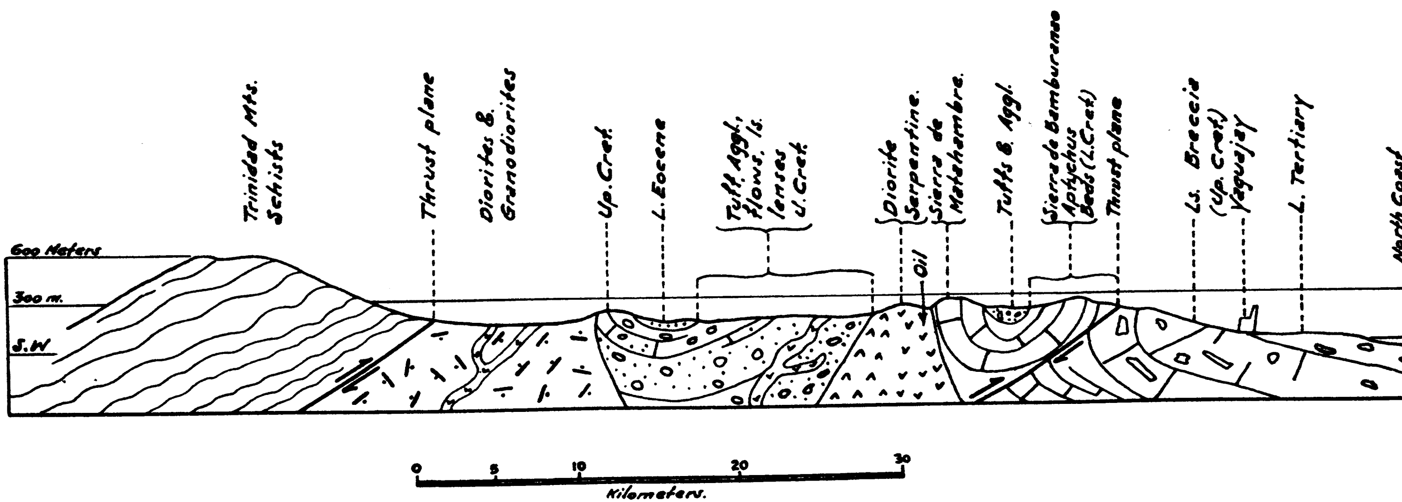


FIG. 5.—Northeast-southwest section (*A–A'*) through eastern Santa Clara Province. Stratigraphically the Limestone breccia of the north phase and the tuffs of the south phase both directly overlie the Aptychus beds. Overthrusting has reversed this position along the north fault and has brought the tuffs of the south phase to within 5 kilometers of the north limestone phase. Late in 1943 a light oil was encountered at shallow depth along the serpentine-Aptychus beds contact in the Sierra de Matahambre.

ceous. There are granites, granodiorites, andesites, rhyolites, basalts, and boulders of sedimentary origin, such as slates, quartzites, schists, and gneisses, whose sources are so hidden that they are but themes for speculation. The boulders in the Cretaceous exposures in the southern or Madruga anticline average somewhat larger than similar boulders in the northern anticline. This has suggested that the mass supplying them lay to the south.[41] Evidence in line with this suggestion is found in the similarity of the Upper Cretaceous shore fauna of Jamaica and Cuba. This in many cases amounts to identity of species. Such similarity in a distinctly shallow-water fauna is difficult to explain if it be assumed that during Cretaceous times Jamaica was separated from Cuba by an expanse of deep water as it is today.

The Upper Cretaceous fauna is southern European in its affinities and distinctly not North American. It is a part of the Caribbean fauna which in turn belongs to the Tethyan or Mediterranean realm of the Old World.

TERTIARY

CAPDEVILA FORMATION, PALEOCENE

The Capdevila formation is a thick series of shales, sandstones, and a few conglomerates which is rather widely distributed in western Habana and eastern Pinar del Río provinces. It weathers to an ochreous brown soil that is rather poor for agriculture.

The Capdevila is folded with the Upper Cretaceous, and in some localities its lithology recalls the Big Boulder bed. Its scanty fauna is confined to Foraminifera and Radiolaria. Some of the Foraminifera are Midway Eocene in age. As pointed out by Brodermann,[42] this is probably a transitional formation. It is well exposed at Capdevila, 10 kilometers south of Habana.[34]

EOCENE

P. J. Bermúdez[44] described the white marls on the University of Habana campus as Lower Eocene and named them the "Universidad de la Habana formation." In its typical exposures in Habana this formation is a white- or cream-colored marl less than 100 feet in thickness.

Deposits ascribed to the Middle Eocene occur in all the provinces of Cuba, the greatest development being in Habana Province in a part of the Madruga anticline known as the Bejucal uplift. Marls, sandstones, and fine-grained conglomerates are the principal elements. These deposits yield excellent foraminiferal faunas representing the Cook Mountain and Mount Selman series of the Claiborne group of the Gulf Coastal Plain. In Oriente Province the thick basaltic flows between Palma Soriano and Santiago de Cuba are Middle Eocene in age. It is possible that some deposits previously referred to the Middle Eocene may be a shallow-water phase of the Upper Eocene.

The Príncipe formation[45] is the principal formation in the Upper Eocene. It is a white- or cream-colored marl that closely resembles the Universidad. It has a large and well-preserved foraminiferal fauna with a few echinoids, echinoid spines, corals, crinoid stems,

[41] R. H. Palmer, "Geology of Habana, Cuba and Vicinity," *Jour. Geol.*, Vol XLII (1934), p. 140.

[42] P. 23 of ftn. 39 (1940).

[43] Palmer, p. 132 of ftn. 41 (1934).

[44] "Estudio micropaleontologico de las formaciónes de las cercanías de la Habana, Cuba," *Mem. Soc. Cubana Hist. Nat.*, Vol. XI (1937), pp. 153–79.

[45] Palmer, p. 132 of ftn. 41 (1934).

and barnacles. At the type locality in Habana City the marls are about 70 feet thick but thicken to the south. Both the Universidad and the Príncipe were deposited in water 100 or more fathoms deep, and both retain much fossil salt. The Príncipe formation weathers to a loose pink or brownish soil that, with sufficient moisture, is rather productive.

OLIGOCENE AND MIOCENE

TINGUARO MARL

The oldest known Oligocene in Cuba is found in the Colon anticline in Matanzas Province. It is a thick series of bluish-gray marls that weathers to a buff or tawny soil which is very compact and impermeable to water. It is poor agricultural land. Good exposures of these marls occur on the lands of the Tinguaro sugar mill and suggested the name for the formation. The marls yield an excellent foraminiferal fauna, part of which was described by D. K. Palmer and P. J. Bermúdez.[46] The Tinguaro marls are believed to be in part the deep-water equivalent of the zone of the large Lepidocyclina that occurs directly above the Eocene in many widely separated localities in Cuba. *Lepidocyclina undosa* and *L. favosa* are two characteristic species of the shallow-water facies. Though these species are referred to the Middle Oligocene or even Miocene in the Gulf States, they occur near the base of the Oligocene in Cuba.

The echinoid fauna of the Upper Eocene and Lower Oligocene, like that of the Upper Cretaceous, has striking Old World affinities. So great are the similarities that many species from Cuba and Egypt are indistinguishable.

The remaining divisions of the Oligocene and Miocene are marls and limestones. For the most part they weather to a residual brown or red clay that ranks high as cane land.

GÜINES LIMESTONE

One of the well-recognized formations of Cuba is the Oligo-Miocene Güines limestone. This is the most widely distributed formation in the island. It is nearly coextensive with the fourth and sixth physiographic provinces. At the surface the formation is a hard, brittle, white or pinkish limestone, in places packed with molluscan and coral molds; a few feet below the surface it is soft, fragmental, and loosely consolidated. Where elevated a few hundred feet, it forms jagged hills such as flank the Habana-Matanzas anticline on the south. The Güines limestone is very porous and easily channeled, with the result that the drainage of this terrain is almost entirely underground. Collapse of cave roofs has formed sinkholes, and in elevated areas a slump or karst topography has resulted. Good examples of this topography are at Jamaica, 25 kilometers southeast of Habana, and in the irregular hills between Limonar and Coliseo in Matanzas Province.

These hills of Güines limestone, though nearly barren of soil, support a dense growth of bushes, vines, and trees. In the underground streams and caves there lives a diversified blind fauna of fishes, spiders, crickets, and shrimps.

The residual soil from the Güines limestone is a red laterite highly prized for agricultural purposes. As stated above, it makes the best cane land in Cuba. The formation of laterite is a tropical or subtropical process or weathering. In the temperate zones the products of weathering are carbonates, silicates, oxides, and hydrous silicates. In

[46] "An Oligocene Forminiferal Fauna from Cuba," *Mem. Soc. Cubana Hist. Nat.*, Vol. X (1936), pp. 227–71.

the tropics the process takes another course or continues to a more complete oxidation, and decomposition of the silicates and the end products are iron and aluminium hydroxides. These insoluble products accumulate as residual soil known as laterite. Laterite in Cuba is derived from both limestone and serpentine. Over large areas in Camagüey and Oriente provinces laterite derived from serpentine has an iron content sufficiently high to make it a valuable iron ore. On the surface of the laterite numerous small pellets of hematite called "perdigones" (shot) have formed. So abundant are these in places that they give a distinctly purplish color to the soil. Under conditions not well understood these perdigones become imbedded in a matrix of limonite, the whole forming a barren soil known as mocarrero.

PLIOCENE AND PLEISTOCENE

Deposits of either of these epochs have not been definitely recognized. Bordering the coast in many places is a narrow collar of hard limestone into which the sea has cut terraces. The city of Habana is located on these marine terraces. Above one of the cliffs is Morro Castle, a well-known landmark of Habana. This limestone contains a few molluscan species and many unidentifiable molluscan and coral molds, but whether the deposits are Pliocene or more recent is not known.

IGNEOUS ROCKS[47]

INTRUSIVES—BASIC

Both intrusive and extrusive igneous rocks occur in Cuba. They in turn are both acid and basic. The basic rocks are for the most part serpentinized intrusions of peridotite, troctolite, or dunite and, with few exceptions, are confined to the northern half of the island. Examples are the serpentine intrusions in the Organos Mountains, the Habana-Matanzas anticline, the Madruga anticline, the Santa Clara overthrust area, central Camagüey, and the disturbed areas in the northern and eastern parts of Oriente Province.

There is no agreement on the time when the serpentine masses were intruded. M. G. Rutten[48] believes them to be pre-Upper Cretaceous. Other unpublished opinions place them in the Upper Cretaceous, Middle Eocene, and even post-Upper Oligocene. The views vary because of the assumption that the serpentines are of the same age, which is not the case. Most of them are intrusive within the Upper Cretaceous terrain, where they have been exposed by erosion. Such are, therefore, post-Upper Cretaceous. Dr. Thomas Thayer, in a personal communication, reports serpentine fragments in the Lower Cretaceous Aptychus beds, implying some pre-Lower Cretaceous serpentine. In western Oriente Province fragments of serpentine occur in the Upper Cretaceous showing a pre-Upper Cretaceous intrusion. The improbability of the Upper Cretaceous age of the serpentine in western Cuba has been pointed out (p. 11). South of Camagüey City the Upper Cretaceous limestone is marmorized near the serpentine contact, indicating a

[47] Schurmann (pp. 339–42 of ftn. 32 [1935]) divides the igneous rocks of Cuba into three classes: (*a*) The intrusions related to the intensive orogeny (Hochorogenese), one of which he places in the Upper Cretaceous. These are the serpentines (syntectonic) and the granodiorite (post-tectonic). (*b*) Older pre-Cretaceous intrusives and extrusives of the Tuff series. These are porphyries and diabases. In contrast to the orogenic intrusions he refers these to geosynclinal types. (*c*) Post-orogenic serpentines, porphyrites, and intrusions of more acid rocks into serpentines and, in southern Oriente, flows of basalt and andesite. These may be Middle Eocene.

[48] P. 13 of ftn. 13 (1936).

later date for that intrusion. In Santa Clara Province a dike is known to cut the Eocene. In Matanzas Province there are serpentine outcrops in an Upper Oligocene limestone terrain. The limestone immediately around the serpentine is marmorized in a zone a few feet wide, while beyond this it is entirely normal. Around the intrusion the Upper Tertiary limestone has dips up to 30°, whereas ordinarily these limestones are flat or, where they border low, post-Oligocene structures, they have dips of 5° or, at most, 8°. In these localities the intrusions appear to be late Tertiary in age. Obviously, there have been several different times of intrusion.

The serpentine-derived soils support a very scant vegetation and are probably the poorest agricultural soils in Cuba. Palma cana (*Sabal parviflora* Bec.), Yuraguana (*Coccothrynax miraguano* Bec.), and Jata de Guanabacoa (*Copernicia macroglossa* C. Wendl.), while not entirely confined to the serpentine, are characteristic of it. However, the lateritic soils derived from iron-rich serpentine support valuable pine forests. Analysis of the soils from the iron-rich and the iron-poor serpentines show a notable percentage of potash in the former and but traces in the latter. The presence of this essential ingredient in the iron-rich serpentine may in part account for its heavy pine growths.

The solid stand of pines is a notable exception in tropical forests. In general, a tropical forest is an indiscriminate mixture of numerous species. Except in high altitudes where temperate conditions prevail, solid stands, like an oak or maple or beech forest of the north, do not occur. It seems strange that these low hills should supply the proper conditions for the pines. The rule seems to be that in the tropics abundant species with few individuals comprise the forests; in the north the reverse prevails. Curiously, the same rule applies to marine molluscan faunas.

The serpentines of Cuba are, however, of economic importance for their local mineral content. The iron-ore deposits that accumulate in the residual laterite from serpentine have already been mentioned. The iron deposits of Mayari and Moa districts in Oriente, in San Felipe in Camagüey, and on Loma Cajalbana in Pinar del Río are examples. The Mayari deposits contain about 1 per cent nickel, which is at present (1944) being exploited. Valuable deposits of chrome also occur in serpentine. They were originally magmatic segregations in the dunite or peridotite, the parent-rocks of the serpentine. There are a number of these deposits scattered over Cuba, the most important being in Camagüey and northern Oriente provinces.

In addition to the serpentine masses, there are many smaller intrusions of diorite, gabbro, and a few of granite and granodiorite in the northern half of the island. In the eastern part of Oriente altered intrusions associated with serpentines form the core of the mountains. Around this core and dipping away from it is a collar of sediments. The intrusions are pre-Tertiary, and the attitude of the Tertiary sediments is due to the late Tertiary folding.[49]

INTRUSIVES—ACID

The southern half of the island has fewer intrusions. On the north side of the Trinidad Mountains in southern Santa Clara Province there are two large granite and granodiorite intrusions whose

[49] For the data and comments on the geology of eastern Oriente I am indebted to Dr. Thomas Thayer.

weathered debris produces a loose, highly siliceous soil similar to that of the Cayetano and, like that soil produces high-grade tobacco. Acid intrusions border the large serpentine mass on the south in southeastern Camagüey Province.

In western Oriente there is a large intrusion of granodiorite and granite occupying an area of 400 square kilometers. In central Camagüey granodiorite and granite are exposed in patches over an area of 250 square kilometers near Florida. In the latter area an attempt was made to remove a supposed boulder, but the project was abandoned when 6 or 8 feet of excavating revealed that the boulder became larger at depth. Instead of being a boulder, it was in reality a knob from a large mass of granodiorite lying buried under the surrounding sediments. The excavation was made in the Habana formation chalk and marl, and these showed no trace of metamorphism along the contact with the granodiorite, indicating the latter to be pre-Upper Cretaceous. A similar granodiorite occurs at Ciego de Avila in western Camagüey, and there are like patches in eastern Camagüey.

Field evidence indicates that the granodiorite intrusions in central and eastern Cuba are approximately the same age. The data in the preceding paragraph clearly show that the Camagüey granodiorite precedes the Habana formation. What precedes the granodiorite, however, is not so clear. Thiadens states: "The contact phenomena in the Tuff Series near the contact with the diorite prove that the Tuff Formation is older than the diorite."[50] L. Rutten[51] states that the granodiorites are post-Tuffs and probably pre-Emscher (Lower Senonian). Both calculations are based on the assumption that the tuffs are Cenomanian-Turonian-Emscher, which, as has been pointed out, is true only of parts of the Tuff series. From the data at hand the only statement that can be definitely made as to the age of the granodiorites is that they are later than some of the tuffs and are older than the Habana formation.

It will be noted that the intrusive rocks in the southern half of the island do not extend west of about the central meridian and that they are acidic types. In contrast, the intrusions in the northern half occur throughout the length of the island and are basic.

[50] P. 12 of ftn. 23 (1937).

[51] "The Age of the Quartzdiorite and Granodiorite Rocks in the West Indies," *Geologie & Mijnbouw*, Jaargang No. 5 (1939), p. 129.

EXTRUSIVES

Extrusive rocks occupy a very small portion of the Cuban terrain. Only two late Pleistocene or Recent unburied flows of any importance are known. One is a comparatively recent flow that covers an area of approximately 25 square kilometers in central Santa Clara Province. The other is an andesitic flow of some 200 square kilometers in Oriente and adjacent parts of Camagüey. In western Camagüey there are two small basaltic cones.

The Cretaceous igneous activity has already been mentioned, and the character of the ejected material and its general distribution discussed. The Cretaceous tuffs, though occupying but a small percentage of the area of Cuba, do in the aggregate amount to many thousand hectares. The tuff-derived soils are for the most part poor and are avoided by agriculture.

There was minor Eocene volcanic activity in the provinces of Matanzas and Oriente, but it was of little importance except for the thick basaltic series that

forms the greater part of the Sierra Maestra in southern Oriente. Taber[52] estimates the Cobre formation, which is largely made up of these flows, to be over 4,500 meters, and "may be as much as 6,000 meters," in thickness.

No Oligocene or Miocene volcanic rocks are known in Cuba.

STRUCTURE

No statement can be formulated to cover the general structure of Cuba. Past attempts to generalize have given impressions that are erroneous. The position of Cuba on the margin of the great Tethyan belt has subjected it to tectonic forces operating both on the American continent and in the Tethyian zone. This, together with its size (1,200 kilometers long), precludes any possibility of a unit structure throughout the entire island. It has, on the contrary, been broken into blocks somewhat smaller than those on the continent which constitute major structural units. An acquaintance with these major units is sufficient to form an idea of the general anatomy of the island. An account of the minor structures would confuse rather than clarify the general picture. The account will therefore be confined to the major structures as far as known, starting at the western end of the island. These structures are located on the map (Fig. 1).

Occupying nearly the entire length of Pinar del Río Province in a strip 4–10 miles wide is the Pinar overthrust. Here the Jurassic and the Viñales limestone that are normally below the Cayetano are thrust southeastward over this formation (Figs. 2 and 3). The dissected, overthrust sheet of Viñales limestone forms the greater part of the Organos Mountains. The overthrusting was more intense in the eastern end of the range, where it was of the order of 10 miles, diminishing toward the southwest and, at the end of the range, dying out.

The hard Viñales and Jurassic limestones advancing over the soft Cayetano plowed deeply and crushed and contorted the shales before them. After the movement ceased, the front of the limestone sheet retreated as a result of erosion, leaving a valley between the limestone mass on the north and the wall or scar in the Cayetano on the south gouged by the overthrust front. North of Pinar del Río City on the Viñales road the Cayetano wall or overthrust scar is 150 feet high, and the shale beds in the base are crushed and contorted (Fig. 3).

The overthrust was not the simple case of a limestone sheet thrust upward to the surface along a straight line and then southward over the Cayetano. Instead, the overthrust mass broke into segments that came to the surface en echelon along a general line. Examples are the Pan de Guajaibon, about 2 miles north of the general line, and the Sierra Quemado, which is 5 miles south of it. Nor is there everywhere only a single overthrust in a given segment. There are cases of two and even three imbrications. A notable example is the segment containing the two parallel ridges, Sierra del Ancon on the north and Sierra de Viñales on the south. The overthrust apparently started along the southern ridge. This was followed by a second slice on the north which passed over the first. Erosion has reduced the two, leaving the front of the second mass as a steep cliff above the first, whose gentler incline has the appearance of a talus slope (Figs. 3 and 4).

The overthrust structure has been complicated by contemporaneous and possibly minor subsequent folding. This has left the Organos Mountain area

[52] P. 577 of ftn. 2 (1934).

folded in a broad anticlinal structure along the central part of which lies the Pinar overthrust (Fig. 3). The entire column from the Jurassic to the Miocene participated in this folding. The name "Pinar anticline" is proposed for this extensive structure.

The Cayetano, very probably, and possibly the Habana formation capped the Viñales limestone during the overthrusting, but they have since been eroded from the overthrust sheet, leaving no trace of their former presence. Erosion has dissected the limestone sheet and cut it into blocks with nearly perpendicular sides varying in size from small hills to masses of several square kilometers often 500 feet high. These isolated blocks are the "mogotes," the most striking feature of the Viñales landscape, as already stated (Fig. 2).

The pressure causing the overthrusting was from the northwest (see Fig. 1). The configuration of Cuba suggests a possible correlation between several features. Pinar del Río Province has a bearing of N. 70° E., and the adjacent part of the island an east-west bearing, a difference of 20° (see Fig. 1). The 1,000-fathom line of the Gulf of Mexico makes a deep re-entrant off the north coast of Pinar del Río. Southeast of the province is the Isle of Pines, and between it and the mainland are shoals. This combination of features suggests that pressure from the northwest, using western Habana Province as a pivot, moved the western end of the island to the southeast through an arc of 20°. In the northern half of the Isle of Pines the structural lines lie in a north-south direction. This does not oppose the above suggestion. In the southwest quarter of the island the Sierra la Cañada and Cerros del Monte have a strike of N. 55° W. and in the southeast quarter strike east-west with a dip of 10° to 15° north. This is the general attitude of the schists and phyllites of the area. Late movement from the southwest would explain this marked change in the latter structural lines.[53]

A second major structural unit lies in the provinces of Habana and Matanzas. Here two large anticlines or anticlinoria, with an intervening syncline, are the principal features. These lie in an east-west direction. The northern structure, called the Habana-Matanzas anticline, extends from Matanzas City westward to Habana, a distance of 90 kilometers (55 miles). At Habana the structure passes under the Lower Eocene Capdevila formation and continues into Pinar del Río Province. Along the axis of the structure erosion has removed the Tertiary capping and exposed the underlying Cretaceous in a belt 10 kilometers (6 miles) wide except about midway between Habana and Matanzas, where a narrow saddle of Oligocene remains. On this high saddle the Hershey sugar mill is located. The Cretaceous is composed of soft shales, sandstones, and tuffs that are easily eroded. The Oligo-Miocene, on the other hand, is a hard limestone that resists erosion. The resulting topography is a broad valley with a Cretaceous floor flanked on either side by steep Upper Tertiary cliffs. Prominent portions of this wall of cliffs have specific names, as Escaleras de Jaruco, Sierra Camerones, El Palenque, and Pan de Matanzas. As

[53] This suggestion is in keeping with B. Willis' ("Isthmian Links," *Bull. Geol. Soc. Amer.*, Vol. XLIII [1932], p. 927) theory of sinking basins (in this case the Gulf of Mexico) and rising and expanding borders. Schuchert (pp. 37, 368, 400, 496 of ftn. 24 [1935]) mentions specific application of this theory to the area in question. On the other hand, this theory would recognize the overthrusts toward the north in central Cuba as overthrusts from the Yucatán Basin (Antillean Sea of Suess and Schuchert) of the Caribbean Sea.

already stated, many serpentine intrusions occur along the valley. The east end of the valley is drained by Yumurí River and is known as the valley of the Yumurí. The river has cut through the eastern end of the anticline near Matanzas City, forming the picturesque and well-known Yumurí Gorge.

Paralleling the Habana-Matanzas anticline and lying 20 kilometers to the south is the Madruga anticline named for the town located near the middle of the structure. This structure extends from central Matanzas westward through Habana Province and into Pinar del Río Province nearly to the Organos Mountains, a distance of 150 kilometers (90 miles), where it flattens out and disappears. The folding was less intense in the Madruga anticline than in the structure to the north, with the result that the Cretaceous is exposed in a few places only. Elsewhere the Eocene forms the surface along the axis, and in Pinar del Río, where the structure dies out, only the Oligocene is exposed.

Between the two major anticlines is the Almendares–San Juan syncline, named from the two rivers that drain each end of the structure. Near the eastern end the synclinal axis turns to the northeast around the eastern end of the Habana-Matanzas anticline and passes under the water. This submerged portion forms Matanzas Bay.

Santa Clara Province forms a third geological unit. It is characterized by overturned folds, overthrusts, and intrusions. There are three principal structural features in this province: the Cordillera along the north coast, the Santa Clara intrusion in the central part, and the Trinidad Mountains in the southern part.

There are many minor structures in Santa Clara that are more or less subsidiary to the major features. A description of these, however, belongs to detailed accounts and is not within the scope of this paper.

The Cordillera is a series of overturned folds and overthrusts. The force causing these structures was directed toward the north and northeast. The Middle Eocene is the youngest formation observed to have participated in the folding of the Cordillera. A small overthrust of Lower Cretaceous over Middle Eocene occurs 5 kilometers west of Sagua la Grande. The folding is, therefore, late Middle or early Upper Eocene in age. There are several intrusions along the Cordillera, the largest being in the east end.

The Santa Clara intruded zone, in the central part of the province, is a complicated series of rocks ranging from serpentines to granites. The details of this zone and the relationships of the various intrusions have not been worked out.

The Trinidad Mountains (see Figs. 1 and 6) in the southern part of Santa Clara Province is a thick block of schists overthrust from the south and lying on granite and granodiorite. The mass is 80 kilometers (50 miles) wide by 30 kilometers (20 miles) long in a north-south direction. The granodiorites are exposed along the north side of the mountains and are separated from the schists by a highly sheared fault plane (Fig. 5). The main mass of the schists has a uniform south dip. On both the east and the west ends of the mountains the tuffs, Upper Cretaceous limestone, and Tertiary strike north-south and dip away from the schists (Fig. 6). Between the sediments and the schists are wide shear zones. Farther to the north these formations resume the east-west direction in accordance with the general grain of the country.

Thiadens,[54] in a brief discussion of the Trinidad Mountains, considers the possibility of their being an overthrust mass but believes that "the schist-formation is an autochthonous complex."

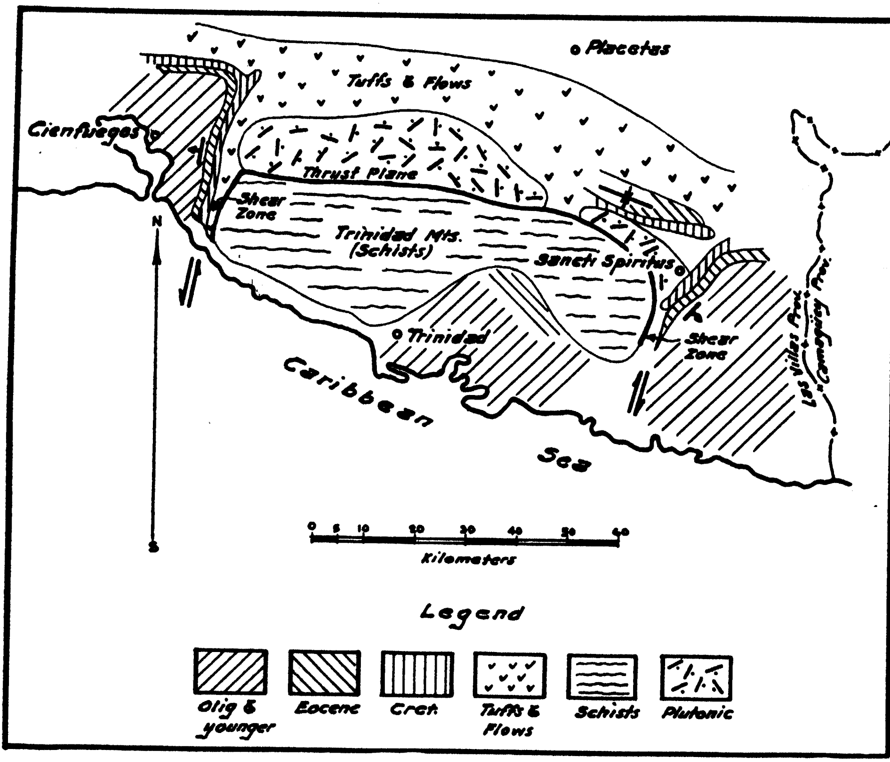


FIG. 6.—Plan of Trinidad overthrust. The Trinidad Mountains are a mass 80 kilometers wide bounded on the east and west by north–south-striking Cretaceous and Tertiary. The mass was pushed northward during late Upper Eocene. North of the mountains earlier movement in late Middle Eocene caused folding in the tuffs, and still farther north in the Cordillera it caused overturning and overthrusting. See Fig. 5.

The overthrust features of central Cuba are suggestive of the Alpine structure of southern Europe with the Trinidad overthrust analogous to the alpine nappes. The Caribbean structures, however, were less intensive and on a much smaller scale. L. Rutten[55] calls attention to the similarity of the Mediterranean and Caribbean regions.

Reference to the map (Fig. 1) suggests a possible correlation between the overthrust features and the topography of the adjoining sea bottoms to the north and south. The 1,000-fathom line on the south side of Cuba makes a detour of 60 kilometers (35 miles) to the north as it passes Santa Clara Province. North of this detour on the mainland are the Trinidad Mountains, followed by a second overthrust zone (Fig. 5). Off the north coast is Cayo Sal Bank. This combination may be viewed as a segment of

[54] P. 48 of ftn. 23 (1937).

[55] "Over de tektonische positie van West Indie," *Natuurwetensch. Tijdschr.* Vol. XVIII, No. 2 (1936).

the surface that was pushed northward, bringing the deep water near the south coast and causing telescoping of the land, a part of which was taken up by the overthrusting on the mainland and a part by the elevations forming the shoals of Cayo Sal Bank. The slight curve in the north coast of Santa Clara and the direction of the overthrusting on the mainland fit into the picture without alteration. According to this suggestion, the Trinidad Mountains were pushed from their former position east of the Isle of Pines northward to their present location, a distance of 50 kilometers (30 miles). Overthrusting on the north side of the province thrust the Aptychus beds over the Upper Cretaceous Limestone breccia and brought the south sandstone-shale-conglomerate-tuff Upper Cretaceous phase in close proximity to the northern limestone phase (see section, Fig. 5). It will be noted that the pressure here was directed to the north and northeast (the *Vergenz* of Stille) in contrast to western Cuba, where it was toward the southeast. In the former case it came from the Caribbean Sea and in the latter from the Gulf of Mexico (see p. 22).

In this connection José Isaac Corral[56] presents a noteworthy theory that is an application of the theory of Continental Drift. His thesis is that the Greater Antilles were formerly adjacent to and a part of Colombia and Venezuela and that during the Upper Miocene they were detached and drifted northward to their present position. This theory may find some support in the overthrusts and overturned folds directed toward the north in Santa Clara Province.

In the southwest corner of Santa Clara Province there is an area outside the overthrust zone. A low, broad, east-west fold known as the Rodas anticline exposes the Cretaceous for a distance of 60 kilometers. Here and in southern Camagüey are the rare places in Cuba where the Upper Cretaceous may be seen lying flat or nearly so.

In Camagüey Province no major structure can be delimited in its entirety. The western half of the province is a plain covered by flat-lying Upper Tertiary limestone. In the north-central part there is an immense structure in which 27,000 feet of Middle to Upper Cretaceous limestone (Jaronú) with a southwest dip has been measured. This structure passes under the Tertiary both to the northwest and to the southeast, and the north flank is under the Bahama Channel. Whether the north side is the limb of an anticline or is faulted off is not known.

In the northern part of western Camagüey there are several structures that bear strong evidence of being salt domes. In general, they have the form of domes; in two places there are large deposits of gypsum of the salt-dome type from which the Tertiary deposits dip. Salt springs occur associated with the gypsum. At present the general area is being investigated by two of the large oil companies.

In eastern Camagüey the most important structure is the large basic intrusion on which Camagüey City is located. This is bordered on the north and on the southeast by rhyolites and diorites and on the south and west by tuffs.

In Oriente Province the major structures are known only in part. The dominant structure in the south is the large and highly complicated monoclinal fold that forms the Sierra Maestra. The basalt and associated tuffs in this fold dip northward from the long system of

[56] "La Union de Cuba con el continente Americano," *Rev. Soc. Cubana de Ingenieros*, Vol. XXXIII (1939), p. 679.

faults that bounds the Bartlett Deep on the north. North of this is the broad syncline of the Cauto Valley. This may be conveniently referred to as the Cauto syncline. This structure extends northward to the intruded zone at Holguin. Along the northeast coast is a folded zone that has been extensively intruded by serpentine. Both the Upper Cretaceous and the Middle Eocene are involved in the folding. Mention was made in the discussion of igneous rocks of the extensive intrusions in the northeast part of Oriente that form large structures surrounded by Tertiary marls and limestones. The prevailing structure of the metamorphics reported in the east end of Oriente is unknown.

HISTORY

The geological history of Cuba does not begin with any definite event or events. The oldest rocks of definitely known age are the Jurassic deposits of Pinar del Río Province. These are marine, and there is no evidence that any land existed at that time anywhere in the area now occupied by Cuba. Except for islands of varying size, there was nothing to suggest the present island until the Middle Miocene. To begin the geological history with events prior to this period is akin to ascribing to its political history events that occurred prior to 1492. The following is rather a history of that part of the Caribbean now occupied by the island of Cuba.

The known geological history of the area starts with the Jurassic. The accounts of events described as occurring in the pre-Cambrian, Paleozoic, and early Mesozoic may be dismissed as fantasies. They have been aptly, though conservatively, described as "contemplative geology . . . . giving a history essentially hypothetical."[57] It is true that the crystalline limestones and schists on the Isle of Pines and in the Trinidad Mountains appear to be old. There is, however, an entire lack of paleontological evidence from which their position in the column can be determined. They may with equal reason be ascribed to the Paleozoic or to the Tertiary.

There were several periods of folding. As far as known, only one, the first, was general throughout the island. The remaining orogenic events were more or less local in character, as is evident from the subsequent account.

JURASSIC

The distribution of the Oxfordian Jurassic indicates marine conditions in western Cuba. It is unknown elsewhere in the island. The thin, limy beds with a bituminous content were deposited in quiet waters at no great distance from land, but whether the land lay to the north or south can only be guessed.

In central Cuba and eastward in Camagüey Province the appearance of Upper Jurassic (Quemado formation) may be due to a general inundation of a large land mass or local embayments in such a mass. The localized occurrences near the north coast in central Cuba suggests the latter alternative. The clastic, calcareous sandstone of the Quemado was apparently deposited along the shore of a transgressive sea. Judging from later events, this land mass probably lay to the south.

In central Cuba the Upper Jurassic Quemado formation passes into the Lower Cretaceous Aptychus beds without a break. No evidence of orogeny between them has been noted.

The account of the late Jurassic orog-

[57] Hayes, Vaughan, and Spencer, p. 30 of ftn. 3 (1938).

eny described by Schuchert[58] was based on a failure to recognize the Pinar overthrust. The supposed unconformity between the Viñales (Aptychus beds) and the underlying younger Cayetano is the overthrust zone between the two.

CRETACEOUS

The local inundations during the Upper Jurassic were followed by complete submergence of at least the northern half of Cuba in the Lower Cretaceous (Aptychus beds). The coarse basal sand followed by the uniformly thinly bedded Aptychus-bearing beds reflect a transgressive sea that covered the five western provinces and probably also Oriente. The presence of Radiolaria and vegetal fragments and the uniformly fine-grained, thin-bedded marls indicate quiet waters at some distance from the shore but not beyond currents bearing debris from land.

A period of intense orogeny followed the deposition and consolidation of the Aptychus beds. This orogeny is unique in two respects: (1) it was the first definitely dated and (2) it was the most intense in the entire geological record of the island. From eastern Pinar del Río to eastern Camagüey badly sheared, contorted, and calcite-veined Aptychus beds are evidence of its intense, far-reaching effects. As far as is known, the activity in Cuba was confined to the northern half of the island. Here structural lines were laid out that were followed by all subsequent orogenies.

This folding event occurred within the North American realm and is of importance. In spite of its intensity and its locality, search for orogenic events on the North American continent comparable in time is vain. South America provides the only possible analogue in the New World. H. Stille states: "The South American Andes afford us on the whole the most striking example of subherzinian (intra-Upper Cretaceous) folding known to the present time."[59] While this pre-Laramide folding of the Aptychus beds cannot be dated except as between their deposition (Neocomian?) and the pre-Maestrichtian, this limit places it within the range of the Andean folding and suggests orogenic relationship between the Greater Antilles and South America. His statement that "weak subherzinian (intra-Upper Cretaceous) movements may likewise exist in Cuba"[60] may well have been prophetic of the discovery of this first great orogenic event in Cuba.[61]

TUFFS

During the deposition of the latest Aptychus beds there was initiated a long period of volcanic activity which continued during the rest of the Cretaceous and to the Middle Eocene. The center of this activity was in central Cuba, where 25,000 feet or more of tuffs, agglomerates, and flows accumulated. The activity extended to all the other provinces but with a lesser degree of intensity and was intermittent, both in space and in time. There are long and wide belts of deposits, long and narrow ones, and small isolated areas, and in many places they are entirely missing. Much or nearly all of Cuba was beneath the sea during this long period of vulcanism, as the numerous intercalated limestone lenses and Radiolaria and Foraminifera within the tuffs bear witness. Cuba, then, was in

[58] Pp. 33 and 517 of ftn. 24 (1935).

[59] "Die Entwicklung des Amerikanischen Kordilleren-systems in Zeit und Raum." Sitzungsber., *Preuss. Akad. Wiss., Phys.-Math. Kl.*, No. 15 (1936), p. 14.

[60] ". . . . schwache subherzynische Bewegungen mögen gleichfalls in Cuba vorliegen."

[61] P. 17 of ftn. 59.

effect a row of volcanic islands not unlike the inner row of volcanic islands in the Lesser Antilles. In time the activity was long ranged. In western Cuba it was Upper Cretaceous, in south-central Cuba from the Aptychus beds through the Upper Cretaceous, and in eastern Cuba it was Middle Eocene. In the latter case it was contemporaneous with similar activity in Haiti.

UPPER CRETACEOUS

The deposition of 34,500 feet of Cayetano consumed a long period of time. No explanation is offered for this formation's being limited to the province of Pinar del Río. Whether the remaining part of Cuba was above water and hence received no sediments or whether it was submerged and received sediments that were subsequently eroded cannot be answered. The high colors of the formation when weathered and the scarcity of fossils suggest estuarine or even fresh-water deposition. Either of these postulates a much larger land mass in this part of the Caribbean than now exists. The limestone lenses with marine fauna may be due to local incursions of the sea. A noteworthy query is the origin of 34,500 feet of Cayetano sediments or the 30,000 cubic kilometers of this formation that has been eroded from Pinar del Río. Cuba could provide no bulk of this order. The sands predominate in the exposures on the north side of the Organos Mountains and shales on the south side, suggesting that transportation was from the north, an area now occupied by the Gulf of Mexico. This is in accord with the interpretation of the Lime gravels, but quite the opposite conclusion was reached in the case of the Big Boulder beds already described and also a part of the Habana formation.

The events occurring during the deposition of the Upper Cretaceous were outlined in part in the description of the Habana formation. The main series of events was, in general, as follows. It has been suggested that possibly most of Cuba was above water while the Cayetano was being laid down. At any rate, except for a few volcanic islands, Cuba was entirely submerged during the deposition of the Habana formation. This is shown by the presence of these marine beds both inland and near the coast in all the provinces of the island. In Pinar del Río, Habana, and Matanzas provinces the succession of Lime gravel, Cone sandstone, and Big Boulder beds indicates a near-by rising land which furnished coarse material. The presence of alveolinellid Foraminifera in Lime gravel pebbles is evidence of Cretaceous exposures in that terrain. The Cone sandstone, with its finer grain, may be interpreted as resulting from a lowering of the topography, possibly due to erosion. The location of the land mass is problematical. The finer-grained Lime gravels in the south anticline, as compared to the north, point to its location on the north. The advent of the Big Boulder bed with an almost complete change in lithology witnesses an equally marked change in the topography and perhaps in the location of the catering land mass. For the reason pointed out, namely, the larger boulders in the south anticline, this land appears to have lain to the south, in the opposite direction from the previous land that furnished the Lime gravels. The north limestone deeper-water phase (Jaronú formation) and the south conglomeratic phase of the Big Boulder bed (see Fig. 1) lend support to this suggestion.

The scattered volcanic activity during the deposition of the Big Boulder bed has already been noted. The tuffs with

marine fauna imply rather small volcanic islands.

It is a curious commentary that, in spite of the orogenic disturbances presumably necessary to produce the marked changes within the Maestrichtian sediments and of the vulcanism in that period, no evidence of any general movement during the Maestrichtian has been observed in Cuba itself.

Attention has been called to the Old World affinities of the Cuban Upper Cretaceous and Middle Tertiary faunas. The shallow-water, shore type of fauna makes the similarity the more striking. The explanation seems to lie either in a migratory route in shallow water along the shores of a Gondwana Land or Isthmian Link, as discussed by Willis[62] between North Africa and the Caribbean or in resorting to the Wegnerian theory of continental drift.

The granodiorite masses in Camagüey and western Oriente provinces, partly buried under unaltered Upper Cretaceous sediments, have been mentioned (p. 20). The evidence there leads to the conclusion that a large island mass was elevated sufficiently above the sea to allow erosion to expose the erstwhile deeply buried plutonic rocks and to reduce the island nearly to base level. The whole island was then lowered to receive the Habana formation sediments. The even larger mass of granodiorite in western Oriente Province appears genetically the same as that in central Cuba. The suggestion comes to mind that the various granodiorite masses are parts of a single mass more than 200 kilometers long extending through Camaguey and into Oriente Province. The existence of this Upper Cretaceous granodiorite island or islands has not heretofore been noted.

[62] P. 929 of ftn. 53 (1932).

ORIGIN OF SEDIMENTS

The south phase of the Habana formation and its equivalent are the only formations whose lithologic content evidences a foreign source. The shales, marls, chalks, and limestones in the various other formations are of material derived from near-by land or from the sea itself.

The source of a large portion of the sands and boulders in the few known Tertiary conglomerates can with considerable confidence be referred to antecedent formations, easily available through natural agencies of transportation. The siliceous content of the Cretaceous Cayetano formation is the logical parent of the sands and shales of the Paleocene Capdevila. The thinly bedded radiolarian limestone and black chert boulders in the Limestone breccia, and the Habana formation in the northern phase, with but little doubt were derived from the contiguous Aptychus beds. Search is in vain, however, for a parent of the highly diversified boulders of the Habana formation above listed (p. 13), nor could any older formation in Cuba have provided the bulk for the 34,500 feet of Cayetano. A source outside of Cuba must be sought. For reasons pointed out, the catering land mass must have lain to the south. The size of this Upper Cretaceous land was considerable. It bordered the entire south coast of Cuba except perhaps the eastern end. It extended an unknown distance to the south but at least as far as Jamaica, judging from the identity of the Upper Cretaceous shore fauna in Cuba and Jamaica. In Curaçao the rudist genera *Vaccinites* and *Durania*, which are common shore genera in Cuba, suggest land connections with northern South America. The presence of *Chiapasella pauciplicata* and other shore species in Cuba and Chiapas

indicate a connection with southern Mexico.

No rudist or molluscan fauna thus far reported from northern Mexico shows any relationship with the Upper Cretaceous fauna of Cuba. The same comment applies to Hispaniola and Puerto Rico. From the present data this land mass was therefore bounded in part on the north by Cuba and on the east by a line passing between Jamaica and Hispaniola and as far south at least as Jamaica and possibly to the north coast of South America. The lack of data on Upper Cretaceous history of Central America prevents any attempt to locate this line on the west except that it included Chiapas in southern Mexico.

PALEOCENE

At the close of the Cretaceous or the beginning of the Paleocene there was mild orogeny on the site of the future Organos Mountains in western Cuba. This is the first known folding in the western end of Cuba. It took the form of a low elevation on which the Cayetano formation was raised above sea-level and exposed to erosion. The very siliceous content of the Cayetano furnished the sands and shales of the Paleocene Capdevila that are exposed on both flanks and east of the mountains. At a distance from the mountains, this formation lies upon the Habana formation with no indication of unconformity between them. Evidently the elevation was not sufficient to expose the Aptychus beds or their western equivalent, the Viñales formation, as no trace of limestone fragments has ever been reported from the Capdevila near the Organos Mountains. In passing, it should be emphasized that this is strong evidence supporting the contention that the Aptychus beds lie below the Cayetano. It was not yet exposed to erosion while the Capdevila was being deposited.

The Capdevila shales, sandstones, and rare conglomerates indicate near-shore conditions with no accentuated topography in the adjoining land mass. Near Capdevila, the type locality, ripple marks show actual strand conditions. A part of Cuba was at that time above water.

EOCENE

In Habana Province the shallow shore-deposited Capdevila is followed successively by brown and then white marls of the Universidad formation. The latter carry an abundant fauna of pelagic Foraminifera and Radiolaria bearing witness of a rapid submergence of considerable magnitude.

The later part of the Middle Eocene witnessed the second great period of orogeny. In Pinar del Río the presence of large boulders of both Upper and Lower Cretaceous rocks in early, rather steeply dipping, Upper Eocene, shows that the Organos Mountains had been elevated above the sea and were being subjected to rapid erosion. It is probable that the pivotal movement of western Cuba from an east-west to a southwest direction accompanied by the beginning of the Pinar overthrust occurred at this time.

This orogeny is the first of any great magnitude that western Cuba experienced. There is no evidence of the late Jurassic orogeny that produced the "ancestral Organos Mountains" mentioned by Schuchert.[63]

West of central Cuba the Habana-Matanzas and the Madruga anticlines were blocked out. At this time occurred the extensive overthrusting in northern Santa Clara when the Lower Cretac

[63] P. 496 of ftn. 24 (1935).

was thrust over the Upper Cretaceous and the Middle Eocene, the south phase of the Upper Cretaceous pushed northward in close proximity to the northern phase, and the Cordillera, in general, further elevated (Fig. 5). In Camagüey Province the 27,000 feet of Upper Cretaceous Jaronú limestone was folded. It seems probable that this orogenic activity extended to Oriente Province and continued eastward into Haiti and the Dominican Republic. In eastern Cuba there was extensive igneous activity, the record of which is preserved in the basalts of Sierra Maestra. The pillow structure of these basalts and the interbedded limestones are evidence that they were deposited under water.

In addition to the Organos Mountains, land of considerable elevation undoubtedly appeared throughout the areas of the structural features mentioned. The resulting group of elongated islands with east-west axes assumed a form roughly outlining Cuba as now known. This great period of orogeny corresponds in part to the Laramide revolution on the North American continent.

The widespread distribution of the deep-water Upper Eocene indicates a general subsidence during which probably the greater part of Cuba was again flooded. In Habana City the Upper Eocene is a pure-white marl which is for the most part massive, carries w[illegible] fossil salt, and is characterized by a distinctly deep, open-water foraminiferal fauna. Land debris is entirely absent. The foraminiferal assemblage of the marls, according to D. K. Palmer,[4] indicates a depth of water of at least 600 feet. This is about at the edge of or somewhat beyond the continental shelf. In Matanzas, Santa Clara, and Camagüey provinces similar deposits indicate similar conditions.

The Trinidad Mountain overthrust is the only tectonic event noted in Cuba in the Upper Eocene. This is dated on rather good evidence. On the flanks of the overthrust area the Cretaceous is a limestone in which no schist boulders have been found. Above the Cretaceous the Middle Eocene and the lowest Upper Eocene are fine-grained, white chalk. At the top of the Upper Eocene a conglomerate appears with boulders of Trinidad schists. Evidently the schists were not exposed before this time. Though the Trinidad overthrust and the overthrust zone in the northern part of the island both resulted from the same forces of compression, the evidence indicates that the two movements were not contemporaneous but that the one in the north was the earlier.

Widely distributed clastic sediments in the Lower Oligocene record a renewed elevation. These occur in all the provinces except Matanzas (see p. 17), where the Lower Oligocene is a deep-water marl. The greater part of Cuba was elevated, and low islands were exposed which provided the small amount of land-derived material that enters into the composition of these sediments. The wide distribution of the Lower Oligocene marine deposits indicates that the islands were not of great extent.

During the Middle Oligocene, Cuba experienced a mild deformation during which the various formations flanking the structures were elevated. Erosion beveled these exposed areas and provided the diversified surface on which the Güines limestone was deposited. Submergence of that surface took place in the transitional Oligo-Miocene and Lower Miocene and was one of the most

[4] "The Occurrence of Fossil Radiolaria in Cuba," [illegible]c. *Cubana Hist. Nat.*, Vol. VIII (1934), p.

extensive that Cuba has experienced. In the west, islands probably persisted on the site of the present Organos Mountains. All of Habana and Matanzas provinces were open ocean. A few islands marked the Cordillera in Santa Clara, and shallow water covered most of the remaining parts of the province. The sea crossed western Camagüey and covered the southern part of the province, extending eastward into the Cauto Valley of Oriente Province. The deposits laid down during this widespread marine incursion are the Güines limestone. This is the most extensive and most unconformable formation in Cuba. Normally it rests on the Cojimar formation (Upper Oligocene), but it is transgressive on various stages of the Oligocene, on the Eocene, and even on the Upper Cretaceous. During this period there was nothing that suggested the Cuba of the present.

Late in the Lower Miocene there occurred another mild orogeny. Quantitatively the movements were small, but, as shallow water covered the deposits involved, the effects were striking. Cuba emerged and for the first time assumed approximately its present outline. The old structures were rejuvenated, and further folding occurred along the Habana-Matanzas and Madruga anticlines. The latter presented an aspect very different from the present. Instead of the broad valleys that now mark these features the Güines limestone covered them in the form of a low swell. Post-Miocene erosion has removed the Güines from the higher elevations and has exposed the softer Eocene and Cretaceous marls, shales, and sandstones, leaving the line of cliffs retreating down the flanks of the two anticlines as stated in the discussion of the third physiographic unit and the second structural unit. In Santa Clara Province the Cordillera was further folded. In Camagüey and Oriente provinces the history is not known in detail, but post-Miocene elevation and minor folding have been recognized.

Deposits subsequent to the Lower Miocene are rare and confined to the coast or inlets of the coast. Examples are the Middle Miocene La Cruz marls in Santiago de Cuba and Manzanillo in Oriente Province. In Matanzas Province the Middle Miocene is exposed on the Canímar River, in a narrow belt extending westward from Matanzas City, and in the gorge of Yumurí River. In the last two localities the inclination of the Middle Miocene beds is greater than can be attributed to initial dip. As they are on the flank of the Habana-Matanzas anticline, it is evident that post-Middle Miocene folding has taken place in this structure.

During the Pleistocene and Recent, vertical movements of considerable magnitude are recorded in the terraces and in sand-choked river channels. Hayes, Vaughan, and Spencer[65] record terraces in widely separated parts of the island. In each locality there are from three to five. The highest are, in Habana, 200 feet; in Matanzas, 300 feet; in Gibara on the north coast of Oriente Province, 150–80 feet; in Manzanillo, 200 feet; and in Santiago de Cuba, 280 feet; while Point Maisí at the extreme eastern end of Cuba is estimated to be 600 feet in height. These wave-cut terraces indicate wide marine incursions.

The drowned valleys that form the harbors of Cuba and the sand-choked river channels are evidence of former elevations of the land or lowering of the sea. In the channel to Habana Harbor that cuts through hard limestone there is 100 feet of sand; in the Almendares

[65] P. 18 of ftn. 3 (1938).

River channel on the west edge of Habana there is 69 feet, and in the valley of San Juan River in Santiago de Cuba, Hayes, Vaughan, and Spencer[66] noted rounded boulders 70 feet below sea-level. Whether there was actual elevation or a lowering of the sea, possibly as the result of the withdrawal of water required in the accumulation of the Pleistocene ice caps, cannot be answered.

Wave-cut terraces above 35 feet elevation have not been observed in Santa Clara or Camagüey provinces. However, there is striking evidence of folding accompanied by elevation in the Cordillera in eastern Santa Clara. The Jatibonico del Norte River crosses a wide limestone ridge in the Cordillera through a natural tunnel. Above the tunnel and passing over the ridge 200 feet above the entrance to the tunnel is the former clear-cut channel with boulders brought from the igneous terrain in the upper reaches of the basin. Salvador and Sarah Massip[67] have also observed this feature. Along the north coast of Camagüey Province a syncline is forming between the mainland and the outer row of islands. Such phenomena as drowned drainage courses, a warped, oxidized surface that is submerged, and drowned vegetation furnish evidence both of the structure and of its recent advent.

Taber[68] considers the faulting of the Bartlett Deep and the rising of the Sierra Maestra as Pleistocene events.

The clear evidence of recent movements from the eastern tip of Cuba westward to Habana prompts the suggestion that the Pinar overthrust in the western end may have been active at least during the Pleistocene or Recent. It is recognized that earlier orogeny had left local high hills or mountains during the Lower and Upper Eocene. The boulder content of conglomerates of that age bear witness of this. Similar features continued with or without interruption into the Oligocene and probably into the Miocene. The thrusting, however, at least in its final stages, occurred at a later time. The deep and very steep drainage courses in the soft Cayetano shales are the result of recent elevation. These shales border the overthrust sheet on the north and participated in the movement of the overthrust. On the north side of the overthrust sulphide ores occur on the surface of the Cayetano shales, indicating erosion of such rapidity that oxidized products could not accumulate. In the tropics this implies very recent elevation. The limestone mogotes, though subject to acid solutions from the dense growth of vegetation, retain a young topography (Fig. 2). A striking example of recently formed topography is the scar, earlier described (Fig. 3 and p. 21), that marks the south front of the overthrust. Though the terrain is the soft Cayetano shales, the face of the scar is still steep and scarcely gullied, and the top or crest has no more than started to retreat. The relative recency of the uplift of these mountains is well shown in the north coastal plain. The portion of the plain lying north of the mountains has a youthful topography with sharp ridges and active streams. On the other hand, the adjacent portion to the east and beyond the end of the mountains has a mature topography with low, rounded hills and sluggish water courses.

[66] *Ibid.*, p. 34.

[67] *Introducción a la geografía de Cuba* (Habana, 1942), p. 95.

[68] P. 597 of ftn. 2 (1934).

BIOLOGICAL EFFECTS OF GEOLOGY

The vicarious events in the geological history of Cuba have had a vital effect on the biological content of the island.

It is only through such favorable events as the appearance of land that terrestrial or marine shore life was even possible. To a satisfactory understanding of the living fauna and flora as well as the fossil, the general geological background is of material assistance.

Attention has been called to the fact that, with the possible exception of portions of the Cayetano formation, the sediments of Cuba are entirely marine. The absence of terrestrial deposits effectually answers the often-repeated question of why no land-vertebrate fauna is ever found in the Cretaceous or Tertiary of Cuba. The only vertebrate remains are a few Pleistocene forms found in cave deposits near Ciego Montero and Mayajigua in Santa Clara Province.

The striking feature of the Cuban mammalian fauna is its scantiness. A list of the living and fossil mammals of Cuba has been compiled by Dr. G. Aguayo of the University of Habana. This comprises 59 species, and his analysis shows that 3 insectivores, 2 carnivores, 5 edentates, and 9 rodents, or 32 per cent, make up the entire list of strictly land mammals. The remaining 40 species, or 68 per cent, are bats, cetaceans, and a sirenian. To the first group, a passage across water is a fortuitous event and accomplished but rarely. Sufficient time has not elapsed for this group to land more than a few species. The flying and swimming mammals, however, quickly populated the island and adjacent waters, and the species soon reached a maximum.

Three factors contributed to the paucity of the strictly land-mammal fauna. First, the shortness of the time available for its development. As pointed out earlier, Cuba as a large island such as it is today did not exist until after the Middle or Upper Miocene. The small islands that marked its location did not afford either space or variety of habitats for an abundant and diversified mammalian fauna.

The second factor is one of both space and time. The source of the Cuban mammalian fauna appears to be South America. Two migration routes have been suggested. One, over former land connections through Central America and by way of Honduras, Jamaica, and Hispaniola to Cuba. The other, via the Lesser Antilles, Puerto Rico, and Hispaniola, and, finally, Cuba. In either case, only the tapering ends of the migrations reached Cuba in the limited time available. This was a factor in the low census of Cuban mammals.

The third factor, equally important and possibly the most important, is the Pleistocene or recent submergence of Cuba. The terraces along the north coast, and also those on the south coast of Oriente Province, evidence a Pleistocene or later repetition of the Miocene marine inundation. They indicate that the submergence was of the order of 300 feet or more. This submergence, coupled with the fact that many areas were at that time at a relatively much lower elevation than at present (see pp. 32–33), killed off the greater part of the terrestrial life. Except for the descendants of those few types that found refuge on the scattered Pleistocene islands and possessed the hardihood to meet the fierce competition in those crowded quarters, the present fauna as well as the flora of Cuba are post-Pleistocene arrivals.

SOME REVISIONS OF THE LATE CENOZOIC STRATIGRAPHY OF THE SOUTHERN OREGON COAST[1]

EWART M. BALDWIN
Oregon Department of Geology and Mineral Industries, Portland

ABSTRACT

The Elk River beds of the Cape Blanco region, as currently referred to, include terrace deposits of late to post-Pleistocene age and beds of Middle Pliocene age which unconformably underlie them. The writer proposes to restrict the name "Elk River beds" to the terrace deposits in accordance with Diller's original definition and further proposes the name "Port Orford formation" for the beds of Middle Pliocene age, as yet bearing no formational name. The name "Coquille formation" is proposed for estuarine deposits which unconformably underlie the Elk River beds (restricted) just north of the Coquille River mouth. The Coquille formation is tentatively correlated with a part of the Taholah formation of Washington. Two stages of marked elevation above sea-level were recognized, the latter followed by submergence of the coast line to its present position. The formation of the late to post-Pleistocene terrace was accompanied and followed by gentle warping.

INTRODUCTION

Recent studies by the writer along the coast of southern Oregon in 1943 and 1944, while mapping with the State Department of Geology and Mineral Industries, have indicated a need for revision and clarification of the Pleistocene and some Pliocene stratigraphic units exposed in the sea cliffs. A similar study of the Pleistocene stratigraphy of the Olympic coast of Washington was undertaken by the writer in 1938[2] and led to the following conclusions. Beds which had been generally assigned to the Pleistocene are separated by an unconformity; unconsolidated terrace deposits overlie older estuarine sediments. Truncation of the estuarine sediments by the sea formed a wave-cut platform upon which the terrace deposits later were laid. The estuarine sediments are confined to channels cut in the Tertiary rock during a former stage of emergence. The recent studies show that similar stratigraphic units are present along the southern Oregon coast.

DESCRIPTION OF THE TERRACE

Remnants of a prominent marine terrace are well displayed at Cape Arago and Cape Blanco along the southern Oregon coast. The terrace approximates sea-level between Heceta Head and Coos Bay, but south of Coos Bay it rises to as much as 225 feet at Cape Blanco. The wave-cut platform upon which the terrace sands and gravels rest truncates folded Tertiary and pre-Tertiary sediments and igneous rock as well as older Pleistocene estuarine deposits. In some places alluvial-fan material is spread over the terrace top where it borders highlands. The terrace is poorly developed on igneous or pre-Tertiary rock, and it is therefore missing along much of the coast south of Port Orford and is barely visible at Heceta Head. The terrace is several miles wide where the Tertiary sediments border the present shore line, although much of the former surface has been recently cut away by the sea.

[1] Published with permission of the director of the Oregon Department of Geology and Mineral Industries.

[2] "Late Cenozoic Diastrophism along the Olympic Coast, Washington" (unpublished Master's thesis, Washington State College, 1939).

Streams are entrenched where they cross the terrace, but the intervening surface presents the initial slope marred only by bars, dunes, and small areas of fanglomerate. Good sections of terrace deposits are exposed along the beach, where they rest upon the wave-cut platform. The unconformity surface, although undulating, is remarkably smooth except near the former shore line, whose location is indicated by wave-rounded and pholad-bored rock. The plane of unconformity is below sea-level between Heceta Head and Coos Bay, but it is exposed in the higher portions of the terrace from South Slough to Port Orford. Some of the range in elevation of the terrace, as well as the platform in the exposed beach section, might be explained by the initial slope of the old sea-bottom and the relative distance from the old shore line. Although some initial slope of the terrace is acknowledged, warping of the coast line is believed to account for most of the range in terrace and platform elevation in the beach section. Along the old strand line, where initial slope cannot be considered a factor, differential elevations afford more definite proof of warping.

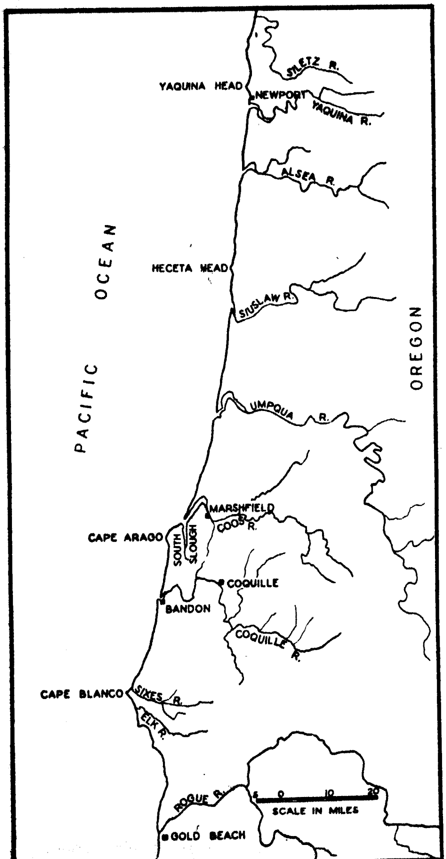


FIG. 1.—Index map of the southwestern Oregon coast.

Although there is a series of higher marine terraces along the Oregon coast, reaching altitudes as high as 1,500 feet, specific reference to a terrace will be confined to the one described, and the others will be referred to collectively.

PREVIOUS WORK

Few writers have dealt with the late Cenozoic stratigraphy and history of the Oregon coast. J. S. Diller[3] described and defined several stratigraphic units during his extensive geologic investigations. W. D. Smith[4] has described many of the coastal features in some detail. J. T. Pardee[5] illustrated many of the coastal sections in his report on beach placers.

[3] "Topographic Development of the Klamath Mountains," *U.S. Geol. Surv. Bull. 196* (1902), pp. 30–31; "Coos Bay, Oregon," *U.S. Geol. Surv. Folio 73* (1901); "Port Orford, Oregon," *U.S. Geol. Surv. Folio 89* (1903).

[4] "Physiography of the Oregon Coast," *Pan-Amer. Geol.*, Vol. LIX (1933), pp. 33–44, 97–114, 190–206, 241–58; Smith and R. E. Fuller, "Oregon Shorelines: A Report of Progress" (Abst.), *Bull. Geol. Soc. Amer.*, Vol. XLI (1930), p. 153; Vol. XLIII (1932), p. 243.

[5] "Beach Placers of the Oregon Coast," *U.S. Geol. Surv. Circ. 8* (1934), pp. 1–41.

More recent studies of the black sands by W. H. Twenhofel[6] and A. B. Griggs[7] have helped explain the origin of the marine terraces. The invertebrate paleontology of Cape Blanco has been recently studied by O. L. Bandy.[8]

STRATIGRAPHY

PORT ORFORD FORMATION

The name "Port Orford formation" is proposed by the writer for Middle Pliocene beds lying unconformably between the Empire and the overlying terrace deposits exposed south of Cape Blanco within the Port Orford quadrangle (Fig. 2). This formation has been included with the terrace deposits under the name "Elk River beds" (or formation),[9] but in reality it is a separate stratigraphic unit.

The basal bed of the Port Orford formation is a buff sand overlain by conglomerate and separated from it by a local unconformity (Fig. 3). Above the conglomerate is a rusty sand which grades upward into a blue-gray argillaceous sand which bears fossiliferous concretions. The top of this member has been truncated by the sea, and the overlying loose gray sand is the basal bed of the late to post-Pleistocene terrace deposits which Diller[10] named the "Elk River beds." He failed to differentiate between the Empire sediments and those referred to here as Port Orford and included both in the Cape Blanco beds.

R. Arnold and H. Hannibal[11] recognized the unconformity between the Empire and the sediments of the Port Orford formation which they believed to be Upper Pliocene, but they included the younger Pliocene beds with the overlying Pleistocene terrace deposits and referred to them as the "Elk River formation."

More detailed work by Bruce Martin[12] confirmed the probable Pliocene age of some of the sediments in Arnold and Hannibal's Elk River formation, but he recognized the Pleistocene age of the upper portion. He concluded that the series was not separated by a stratigraphic break even though the faunas and lithology were sufficiently distinct to allow the series to be separated into two horizons. He evidently failed to recognize the wave-cut surface which truncated the Port Orford sediments and upon which the Elk River terrace deposits rest with apparent conformity but separated by a considerable lapse of time. Martin reported abundant fossils in both the Pliocene and the Pleistocene beds near the mouth of the Elk River. He noted the prominent shell bed at the base of the thin terrace deposits just south of the cape and fossils in the uppermost part of the section near the mouth of the Elk River. The faunas contained a few species not common to both, but the difference in fauna, he concluded, might be caused by the difference in position with regard to the strand line. The time necessary to lay down the terrace

[6] "Origin of the Black Sands of the Coast of Southwest Oregon," *Oregon State Dept. Geol. & Min. Ind. Bull. 24* (1943), pp. 1–25.

[7] "Chromite Sands of the Coast of Southwestern Oregon" (manuscript in preparation, United States Geological Survey).

[8] "Invertebrate Paleontology of Cape Blanco" (unpublished Master's thesis, Oregon State College, 1941).

[9] M. G. Wilmarth, "Lexicon of Geologic Names of the United States," *U.S. Geol. Surv. Bull. 896* (1938), pp. 673–74.

[10] P. 31 of ftn. 3 (1902).

[11] "The Marine Tertiary Stratigraphy of the North Coast of America," *Proc. Amer. Phil. Soc.*, Vol. LII (1913), p. 604.

[12] "The Pliocene of Middle and Northern California," *Calif. Univ. Dept. Geol. Sci. Bull.*, Vol. IX (1916), pp. 219–20.

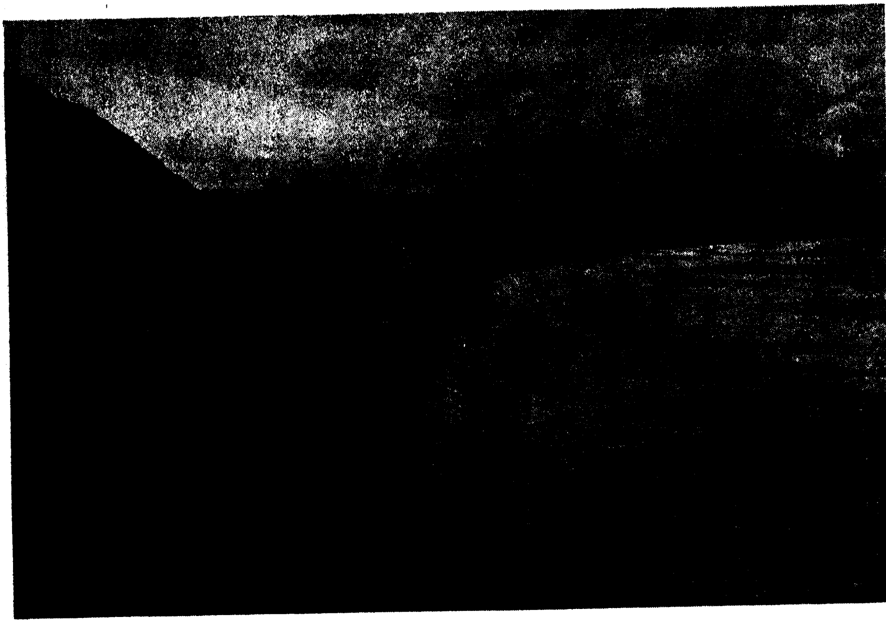


Fig. 2.—Beach section just north of the mouth of Elk River. (1) Elk River beds. (2) Plane of the wave-cut platform. (3) The Port Orford formation. (4) Woody sediments which are tentatively correlated with the Coquille formation (Fig. 6).

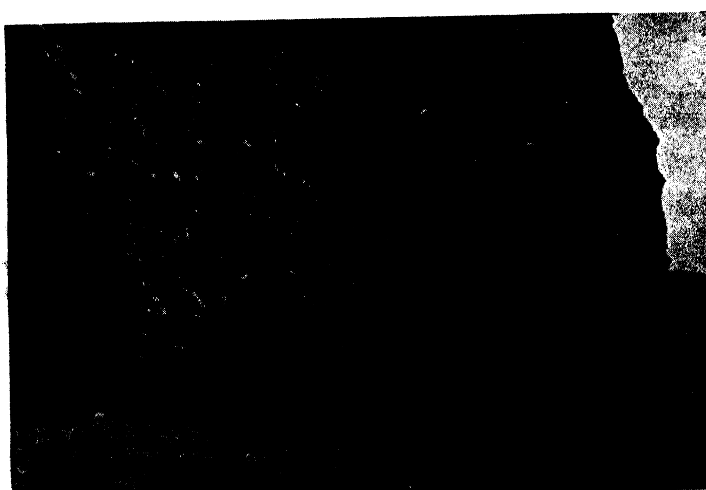


Fig. 3.—Local unconformity in the Port Orford formation

deposits would probably be so short that Martin's explanation for the slight difference in fauna seems acceptable.

W. D. Smith and E. L. Packard[13] mention "two distinct horizons above the Empire (at Cape Blanco), the lower one being presumably the Cape Blanco beds of Diller." However, Diller used the term "Cape Blanco beds" to include both the Empire and the younger Pliocene (Port Orford formation).

cluded all the original Elk River beds (Table 1).

The unconformity is slightly angular at the northern end of the exposed Port Orford sediments, which dip more steeply than the overlying Elk River beds; and the blue-gray argillaceous sand of the Port Orford nearly pinches out, allowing the conglomerate of the Port Orford to be confused with slumped gravels of the terrace deposits.

TABLE 1

HISTORICAL SUMMARY OF THE DIVISIONS PROPOSED FOR THE CAPE BLANCO SECTION

Diller (1902)	Arnold and Hannibal (1913)	Martin (1916)	Bandy (1941)	Baldwin (1945)
Elk River Beds	Elk River Formation	Elk River Beds	Terrace Deposits	Elk River Beds
unconformity			*unconformity*	*unconformity*
Cape Blanco Beds			Elk River Formation	Port Orford Formation
	unconformity	*unconformity*	*unconformity*	*unconformity*
	Empire Formation	Empire Formation	Empire Formation	Empire Formation

Bandy[14] concluded that the sediments here referred to the Port Orford formation were Middle Pliocene instead of Upper Pliocene, as proposed by Arnold and Hannibal and later by Martin. He excluded those terrace deposits which contained the shell bed near the cape and restricted the name "Elk River formation" to the sediments of Middle Pliocene age. This completed the evolution of the name from its original application by Diller to its use for the Middle Pliocene sediments by Bandy, who ex-

[13] "Salient Features of the Geology of Oregon," *Jour. Geol.*, Vol. XXVII (1919), p. 100.

[14] P. 25 of ftn. 8.

COQUILLE FORMATION

The name "Coquille formation" is proposed for the section of gently deformed sediments exposed in the beach section between Whiskey Run and Cut Creek just north of the Coquille River mouth (Table 2) (cf. Figs. 4 and 5).

The measured section was the thickest continuous section, but it does not represent all that is exposed because of intervening areas of slumped Elk River beds. Where the measurement was taken, the dip was as high as 25°, thus exposing 93 feet, but other isolated exposures of the formation which were generally horizontal could not be placed with cer-

tainty in relation to the measured section. The total exposed thickness is probably in excess of 200 feet, and a greater original thickness is implied because the upper portion has been removed by the sea and the base is beneath the sea. At the mouth of Whiskey Run the Coquille formation unconformably overlies the Umpqua formation.

The sediments appear to be estuarine in origin, and, because of their location, they appear to fill a former valley of the Coquille River before its mouth was shifted southward by the present sandspit (Fig. 4).

TABLE 2

MEASURED SECTION OF THE COQUILLE FORMATION

	Feet
Conglomerate	24
Bed containing stumps and peat	2–3
Coarse sand with some grit and logs	18
Sand and woody material above a local unconformity	10
Cross-bedded sand with woody material and an occasional pebble lens	30
Thin-bedded sandy clay	9
	93+

Several feet of clay and peat of similar stratigraphic position are exposed at low tide in the point which separates the south end of South Slough. This is the only place where sediments were observed between the Empire and the Elk River beds.

Sand and clay containing abundant pieces of wood are exposed in the cliff near the natatorium at Newport, Oregon, beneath the wave-cut platform. These sediments are similar in lithology and stratigraphic position to the Coquille formation and are correlated with it.

Sediments which contain abundant woody material are exposed unconformably above the Port Orford formation in the small valley in the beach section midway between Cape Blanco and the mouth of Elk River (Figs. 6 and 2). The stream of this valley is a tributar[illegible] Sixes River drainage and was ap[illegible] beheaded by the advancing s[illegible] bottom of the old valley is onl[illegible] feet below sea-level, so that it is not likely that the Sixes River ever drained through to the sea by this route, in light of previous downcutting exhibited in the Coquille and other river valleys. The sediments are obviously not a part of either the Port Orford or the Elk River beds, and from their stratigraphic position and similar environment of deposition, as suggested by abundant woody material, they may be the equivalent of the Coquille formation.

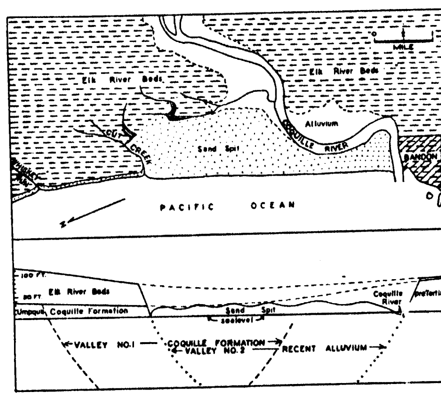


FIG. 4.—Plan view and cross-section of the mouth of the Coquille River.

Similar sediments occupying a[illegible] [illegible]ogous stratigraphic position al[illegible] Washington coast have been st[illegible] by the writer. These sediments, which appear to fill former valleys, are well displayed near the mouths of some of the streams, an example of which is displayed just south of the Hoh River mouth. These sediments have been included by S. L. Glover[15] in his Taholah

[15] "Pleistocene Deformation in the Olympic Coastal Region, Washington," *Northwest Sci.*, Vol. XIV (1940), p. 69.

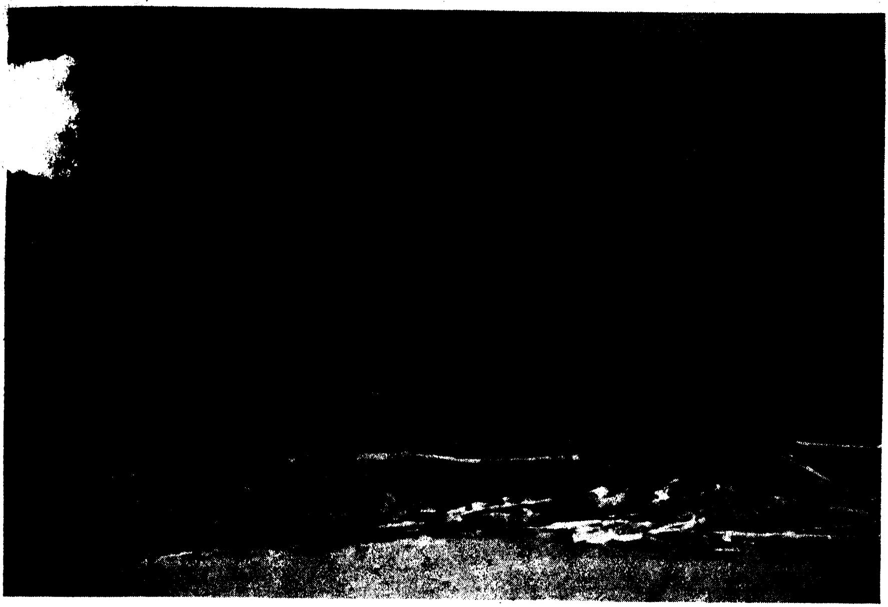


FIG. 5.—Elk River beds lying unconformably above the Coquille formation just south of the mouth of Whiskey Run. The wave-cut platform coincides with vegetation that grows beneath the line of water seepage.

FIG. 6.—Sediments containing intercalated peat and logs which were deposited in a former tributary valley of the Sixes River.

formation along with a much thicker section of sediments in the Taholah-Moclips area, which he described as follows:

> One outstanding example is in the Taholah-Moclips vicinity where a sedimentation phase of the Taholah formation occurs that is quite distinct from the usual type. In this particular area the sands, gravels, and occasional clays were deposited in a subsiding basin, presumably an estuary, and so accumulated to a thickness far greater than is known elsewhere on the coast. Detailed sections of as much as 475 feet have been measured, and the persistent structural attitude of scattered exposures indicates that the total thickness of the formation is in excess of 1,500 feet.

There appears to be no faunal evidence to indicate the age of this section of sediments. However, the Taholah formation seems to include sediments deposited under different conditions: those laid down in erosional valleys and those deposited in a subsiding basin. It would be coincidental if the two were of the same age. Until further work clarifies the limits of the Taholah formation, a correlation of the Coquille formation of Oregon can be tentatively made only with that portion of the Taholah which appears to occupy former stream valleys.

Sediments of the Coquille formation and those believed to be their equivalent are classified as valley fills because they have steep erosional contacts and lie between Tertiary or pre-Tertiary masses beneath the wave-cut platform; they contain coarse-grained to fine-grained beds with local unconformities, cross-bedding, and many pieces of wood; their areal extent is believed to be too small to be explained by subsidence; and they are without known exception near a stream mouth.

If the interpretation that these isolated deposits represent deposition in valleys formed during a prior stage of emergence is correct, sediments of this stage might be expected to be found at the mouths of all the larger streams along this part of the coast. There are at least two factors which prevent this from being true. Erosion during a later stage of emergence has removed the Coquille sediments from the valleys except at the mouth of the Coquille River and in a few other places where a shift of the river's course has preserved them. Then, too, the wave-cut platform is beneath sea-level or buried by extensive sand dunes for many miles in down-warped parts of the coast, so that sediments of this stage, if present, cannot be seen.

The age of the Coquille formation is believed to be very late Pleistocene, but it will be further discussed in the summary of the geologic history.

ELK RIVER BEDS

The Elk River beds, as defined by Diller[16] and used by the writer, consist of the terrace sands and gravels which rest upon the wave-cut platform. Although he did not elaborate as to the presence of or type of unconformity, Diller's diagram of the Cape Blanco region shows this to be true. The sediments range in thickness from 10 feet near the cape to 90 feet just south of the mouth of Elk River. The basal loose gray sand is eroded more rapidly than the underlying Port Orford sediments, thus accentuating the break. Above the basal sands, rusty gravels predominate. Sediments forming this same terrace farther north along the coast of Oregon are more sandy, and north of Coos Bay they are generally re-worked as dunes. The name "Elk River beds" should apply to all the terrace deposits forming this terrace along the Oregon coast.

[16] Pp. 30–31 of ftn. 3 (1902).

Fossils, which Martin[17] concluded were Recent forms, are abundant at the base of the Elk River beds just south of Cape Blanco, and farther south they are intercalated with the thicker terrace deposits. These were studied later by Bandy,[18] who likewise concluded that they were Recent forms and deposited under cold-water conditions.

The upraised marine terraces of the Ventura region of California have been studied by Putnam,[19] who concluded that the higher terraces resulted from diastrophism rather than from fluctuations of sea-level. Faunas on even some of the higher terraces were identified by Grant,[20] who concluded that they suggested a post-Pleistocene age. Bailey[21] indicated that a probable late Pleistocene age for the entire series of marine terraces was to be inferred from their truncation of more steeply dipping beds of Upper Pleistocene age. The relationship between the California and Oregon marine terraces has been pointed out by Gale,[22] who stated:

> The whole western border of the continent seems to have been uplifted gradually as a unit, or the sea level may have subsided eustatically, or both processes may have worked simultaneously. . . . . The youngest series of marine terraces exposed at various places along the coast from San Diego to Oregon seems to be remarkably uniform in general features, though the individual terraces vary in number and height because of minor local warping.

[17] P. 247 of ftn. 12 (1916).

[18] P. 46 of ftn. 8.

[19] "Geomorphology of the Ventura Region, California, *Bull. Geol. Soc. Amer.*, Vol. LIII (1942), p. 752.

[20] *Ibid.*, p. 700.

[21] "Late Pleistocene Coast Range Orogenesis in Southern California," *Bull. Geol. Soc. Amer.*, Vol. LIV (1943), p. 1554.

[22] U. S. Grant, IV, and H. R. Gale, "Pliocene and Pleistocene Mollusca of California," *Mem. San Diego Soc. Nat. Hist.*, Vol. I (1931), p. 64.

Although individual terraces of the Oregon coast may not be correlated with those of the California series, the Elk River beds would appear to be one of the younger—very late to post-Pleistocene in age.

EVIDENCE OF COASTAL WARPING

Warping of some of the younger marine terraces along the California coast has been recognized by Davis,[23] Putnam,[24] and others. Smith[25] has suggested warping along the Oregon coast, and Pardee[26] referred to differential uplift. The youngest of the uplifted marine terraces along the Washington coast has been studied by the writer[27] and the warping plotted.

The datum plane used in measurements was the terrace level at the old strand line, which may be identified by pholad borings and a well-defined notch at the base of the slope back of the terrace, as is so well demonstrated at Yaquina Head. Measurements are based upon hand leveling, triangulation, and United States Geological Survey topographic data. Two datum planes were measured—the top of the terrace and the top of the wave-cut platform—but the evidence for warping is based primarily upon the top of the terrace. The plotted sections in Figure 7, with the exception of that at Cape Blanco, were taken at or near the old strand line and are believed to be correct to within a few feet.

Although the plotted elevations are not evenly spaced along the shore line,

[23] "Glacial Epochs of the Santa Monica Mountains, California," *Bull. Geol. Soc. Amer.*, Vol. XLIV (1933), pp. 1068–69.

[24] P. 752 of ftn. 19 (1942).

[25] P. 101 of ftn. 4 (1933).

[26] P. 31 of ftn. 5 (1934).

[27] Pp. 18–20 of ftn. 2 (1939).

they do indicate warping of the coast line after deposition of the Elk River beds. Thickening and thinning of the terrace deposits suggest warping of the wave-cut platform during terrace deposition, as is shown at Cape Blanco, the highest point on the terrace. The wave-cut platform is 200 feet above sea-level at the cape, but it disappears beneath the sea 3 miles south at the mouth of the Elk River (Fig. 2). The top of the terrace is 225 feet in elevation at the cape and drops to 50 feet at Port Orford, 7 miles to the south. The dip of the underlying

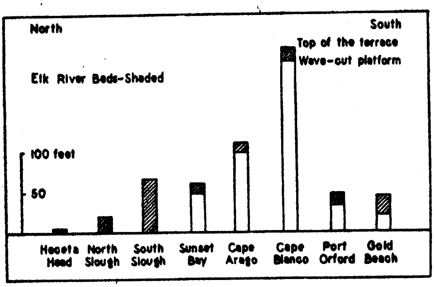


FIG. 7.—Plotted elevations of the top of the terrace and the wave-cut platform.

Port Orford formation is greatest where it overlaps the Empire, probably caused by compaction of the sediments in the basin of deposition; but along the greater part of the section it is generally parallel to that of the warped terrace deposits.

The streams to the north of and including the Coquille River drain relatively low-lying areas of soft Tertiary rock. They have broad flood plains, unfilled bays, and the tidal currents affect the river as much as 35 miles upstream—facts which have led to the conclusion that they have been submerged to a greater extent than rivers to the south of the Coquille River which have a steeper gradient and a narrower valley cut in pre-Tertiary rock. This supposition is not supported by relative elevation of the terrace. An examination of the Rogue River valley indicates that it, too, has a flood plain, but the original valley was narrower, because it was cut in more resistant rock, and the heavily laden stream has aggraded its valley so that the tidal currents do not extend far upstream.

RECENT FAULTING

Small displacements in the terrace deposits are fairly common. One of the larger of the faults, with a 10-foot slip, parallels the steep bedding plane of the Eocene coal bed at Mussel Reef near the Cape Arago lighthouse. This fault, like others in the vicinity, is a small reverse bedding-plane fault.

SUMMARY OF PLIOCENE AND PLEISTOCENE HISTORY

Deposition of the Empire formation in the Coos Bay area started during the latter part of early Pliocene time and continued throughout most of the Middle Pliocene, according to Weaver.[28] Deformation of the Empire beds in the Cape Blanco region preceded the deposition of the Middle Pliocene Port Orford formation, resulting in an angular discordance of 15°–20° and a pronounced erosional break between the two formations.

Deformation of the Port Orford prior to terrace formation seems to have been largely confined to compaction of the sediments, which has caused the beds to dip more steeply away from the Empire near the contact.

From later Pliocene to the Middle Pleistocene was a time of erosion which accompanied and followed the uplift of the Coast Range. Recent work along the California coast, already cited, indicates a late to post-Pleistocene age for even

[28] "Correlation of the Marine Cenozoic Formations of Western North America" (chart), *Bull. Geol. Soc. Amer.*, Vol. LV (1944), opp. p. 596.

some of the higher terraces. Terraces which are found even higher than 1,500 feet point to regional uplift because they seem too high to be accounted for entirely by eustatic changes in sea-level.

Davis[29] has suggested a correlation between terrace formation and stages of glaciation and deglaciation. Although the higher terraces are probably too high to be explained by eustatic changes, two stages of stream cutting beneath present sea-level and the lower terraces may reflect eustatic changes accompanying the last two periods of glaciation during later stages of regional uplift.

If the older of the resultant valleys, the one in which the Coquille formation was deposited, was formed before the higher terraces, the valley would have been filled during the ensuing submergence with resultant changes in stream courses, and evidence of this has not been recognized in the valleys. It does not seem likely that regional trends would be reversed to allow uplift, submergence to at least 1,500 feet, and then uplift again, as demonstrated by the steplike marine terraces. It seems more probable that the older of the two valleys was formed by relative changes in sea-level, probably eustatic in nature, in the latter part of the present regional uplift. Further work in the Puget Sound area will probably help to correlate the history of the Coquille River as outlined with the latest glacial and interglacial stages. It seems probable that the Coquille formation antedates most of the higher terraces and is therefore late to very late Pleistocene in age.

The following stratigraphic column places the late Cenozoic stratigraphic units in their relative order. It must be emphasized that dating of the younger formations is tentative because of the early stage of study of Pleistocene-Recent history along the Pacific Coast (Table 3).

The history of the Coquille River shows the two stages of emergence above present sea-level, and a diagrammatic cross section indicates their probable history (Fig. 4). It is popularly believed that the Coquille River flowed into Coos Bay by way of the "isthmus" near Coaledo which now separates drainage to the Coos and Coquille rivers. Twenhofel[30] indicated that this was the case and that, "when the submergence of the coast began [the submergence prior to the formation of the 1,500 foot terrace and the rest of the terrace series], ... the present valley of the Coquille River into the Bandon area did not then exist."

TABLE 3

Very late Pleistocene to post-Pleistocene	Elk River beds
Late Pleistocene	Coquille formation
Middle Pliocene	Port Orford formation
Upper lower Pliocene and Middle Pliocene	Empire formation

Evidence of two former valleys at the mouth of the Coquille indicates that the river followed approximately its present course even before headward erosion of Beaver Slough, a tributary of the Coquille, and Isthmus Slough, a tributary of the Coos River, formed the pass between the two drainages. The former stages of emergence, shown at the mouth of the Coquille River, would have formed the "isthmus" during stages of downcutting, and therefore it would be unnecessary to postulate deep valley formation prior to regional uplift.

The Coquille formation, deposited in valley No. 1 (see Fig. 4), was some time later truncated by the sea and then covered by a veneer of terrace deposits,

[29] P. 1045 of ftn. 23 (1933).

[30] P. 20 of ftn. 6 (1943).

the Elk River beds. Deposition of the Elk River beds was followed by relative uplift, at which time the present terrace was much higher than now, and valley No. 2 was formed. Later drowning of valley No. 2 has returned the terrace to its present level and filled this valley and its re-entrants as much as 35 miles upstream. Above tidewater, the present valley of the Coquille becomes V-shaped and the gradient increases. Valleys corresponding to No. 2 are filled south of the Coquille River, but to the north, bays and lakes occupy the unfilled portions of these valleys because the streams drain less rugged areas and carry smaller loads.

Diller[31] cited drilling operations which penetrated 200 feet of alluvium without reaching its base, and a projection of the stream and tributary profiles suggests an even greater filling.

Although the present trend of relative movement of sea-level was not determined, the submergence which drowned valleys corresponding to valley No. 2 must be fairly recent because of the unfilled nature of the lakes and bays.

[31] "Coos Bay, Oregon," *U.S. Geol. Surv. Folio 73* (1901), p. 3.

Acknowledgments.—The writer wishes to acknowledge his appreciation for kindly criticism and helpful suggestions while completing this manuscript to Dr. R. L. Lupher, Washington State College; Dr. W. D. Smith, University of Oregon; and Dr. E. L. Packard, Oregon State College; to Mr. John Eliot Allen and Dr. W. D. Lowry, of the Oregon Department of Geology and Mineral Industries; and to F. W. Libbey, director of the Oregon Department of Geology and Mineral Industries, for the opportunity to visit most of the localities mentioned and for permission to publish this paper.

EXAMPLES OF THE INTERPRETATION OF FOLDING

E. SHERBON HILLS
University of Melbourne, Melbourne, Australia

ABSTRACT

Two examples of folded strata are discussed, in which bedding-plane slip is revealed by the displacements of transverse quartz veins from bed to bed. Detailed examination of a specimen containing argillaceous and arenaceous lamellae affords evidence for an interpretation of the internal readjustments by mass flowage, bedding-plane slip, and slip along planes of fracture cleavage, that went on during folding. The fracture cleavage, although parallel to the axial planes of the folds, is believed to have formed obliquely to the *AB* plane of the strain ellipsoid.

INTRODUCTION

The genetic interpretation of minor structures in folded rocks that exhibit cleavage but are otherwise little metamorphosed is not easy, since decisive petrofabric data such as may be obtained from coarsely crystalline tectonites are difficult to obtain. Hans Cloos[1] has, however, demonstrated that under favorable circumstances the nature of certain internal adjustments that have taken place in the folding of such rocks can be determined. In the example he describes, optical evidence of congruous bedding-slip[2] was obtained from slickensides and from the displacements of pretectonic transverse quartz veins from bed to bed.

GOLDEN STAIRS MINE

A similar example, regarded by the present author as comparable with that described by Cloos, occurs at the Golden Stairs Mine, near Melbourne, Victoria, where the section surveyed by J. P. L. Kenny[3] shows several congruous displacements of the auriferous quartz veins from bed to bed on the limb of an anticlinal fold. The rocks are interbedded Silurian sandstones and mudstones, folded into an open anticline with dips, at the Mine, of 20° on the limbs. In the section at Morrisey's Reef a vertical quartz reef crossing the strata shows a succession of offsets along the bedding planes, each of about 2 feet, the movement in the upper bed of any two being toward the crest of the fold. The rocks are seen in the field to be massive, without well-defined stratification planes within the individual beds, which would account for the restriction of bedding-slip to the major bedding planes. The stratal gliding units were the individual beds. In closely folded, laminated Silurian sandstones in the Melbourne area, on the other hand, it has been shown by the author[4] that gliding in folding took place between the thin laminae themselves, as well as on major bedding planes. The thickness of stratal gliding units is particularly important in regard to the position of the neutral surface in neutral surface folding. Clearly, where laminae are the units in gliding, each lamina has its own neutral surface, and the amount of compression and elongation, respectively, below and above this surface is much less in a thin

[1] "Der Gang eine Falte," *Fortschr. der Geol. etc.*, Bd. XI, Hft. 33, pp. 73–88.

[2] In this paper the term "congruous" is applied to such movements of strata as are believed to have resulted from the normal processes of internal readjustment accompanying the formation of individual folds.

[3] "Golden Stairs Mine, Greensborough," *Rec. Geol. Surv. Vict.*, Vol. V, Part II (1936), pp. 222–23.

[4] E. S. Hills, "The Silurian Rocks of the Studley Park District," *Proc. Roy. Soc. Vict.*, Vol. LIII, Part I (1941), pp. 167–91.

lamina than it is in a thick, massive bed. This relationship in turn affects the form assumed by different rocks under similar tectonic conditions and goes far toward accounting for the frequent development of close and acute folds in thin-bedded rocks, as compared with the broader and more open folds of massive strata. The fact that, as will be shown below, laminae only a few millimeters thick may act as stratal gliding units is, therefore, worthy of note.

The implication at the Golden Stairs Mine that the quartz veins were emplaced before the folding was completed is important, since it has been generally assumed that the auriferous quartz reefs in the central Victorian gold fields were introduced after the folding of the Ordovician and Silurian rocks of the region. It would appear that there was some overlap of the metallogenetic epoch and the diastrophic period.

FOLDING IN A LAMINATED ROCK

A large specimen from the gorge of the Mitta Mitta River at Mitta Mitta, Victoria, reveals particularly well, by the displacements of quartz veinlets, the relative movements of stratal gliding units in folding. The rock consists of laminated sandstone and dark, gray-black slate (Fig. 1). The stratification planes in the sandstones are spangled with a phyllitic development of small mica flakes, but otherwise recrystallization is not notable, apart from the normal reconstitution of the slate.

Bedding-slip.—The rock is traversed by several minute quartz veinlets, the offsetting of which in adjacent laminae reveals systematic bedding-slip displacements which are clearly connected with differential gliding or shearing among the laminae during folding (see Figs. 1 and 2). Most of the displacements are congruous with relation to the geometrical form of the folds, but the two veinlets (*I*) show incongruous movement with relation to the adjacent anticline. They may have been injected after the formation of the anticline and have been displaced by bedding-slip during the formation of the small subsidiary syncline on the flank of which they occur. If so, they are of slightly later origin than the group above described. The similarity in trend of the various veinlets suggests that they have all similar histories, yet the widest shows no evidence of displacement by bedding-slip, although it is cracked through parallel to the stratification planes. This indicates that the sandstone laminae were not sufficiently strong to transmit a stress capable of shearing the broad veinlet; the cracks are regarded as shear joints, developed along planes of incipient shearing after some change in the tectonic setting of the rock. The inability of the sandstone laminae to shear this veinlet through may be accounted for, since the veinlet is thicker than any lamina present, and, being composed of a mosaic of crystalline quartz, it would have a greater strength than the granular sandstone, which also contains a little argillaceous material. It is therefore suggested that, even if the folding had progressed further, this relatively thick quartz veinlet would not have been sheared as were the thinner ones, along the planes of lamination of the rock.

Flow in argillaceous material.—The veinlets may still be recognized where they pass into the central dark argillaceous layer (now slate), and it may be seen that there has been no post-vein shearing along the topmost sandy bedding plane (Fig. 1, *B*), since, although mass movement has taken place, the continuity of the veinlets is maintained across this bedding plane. Furthermore, there

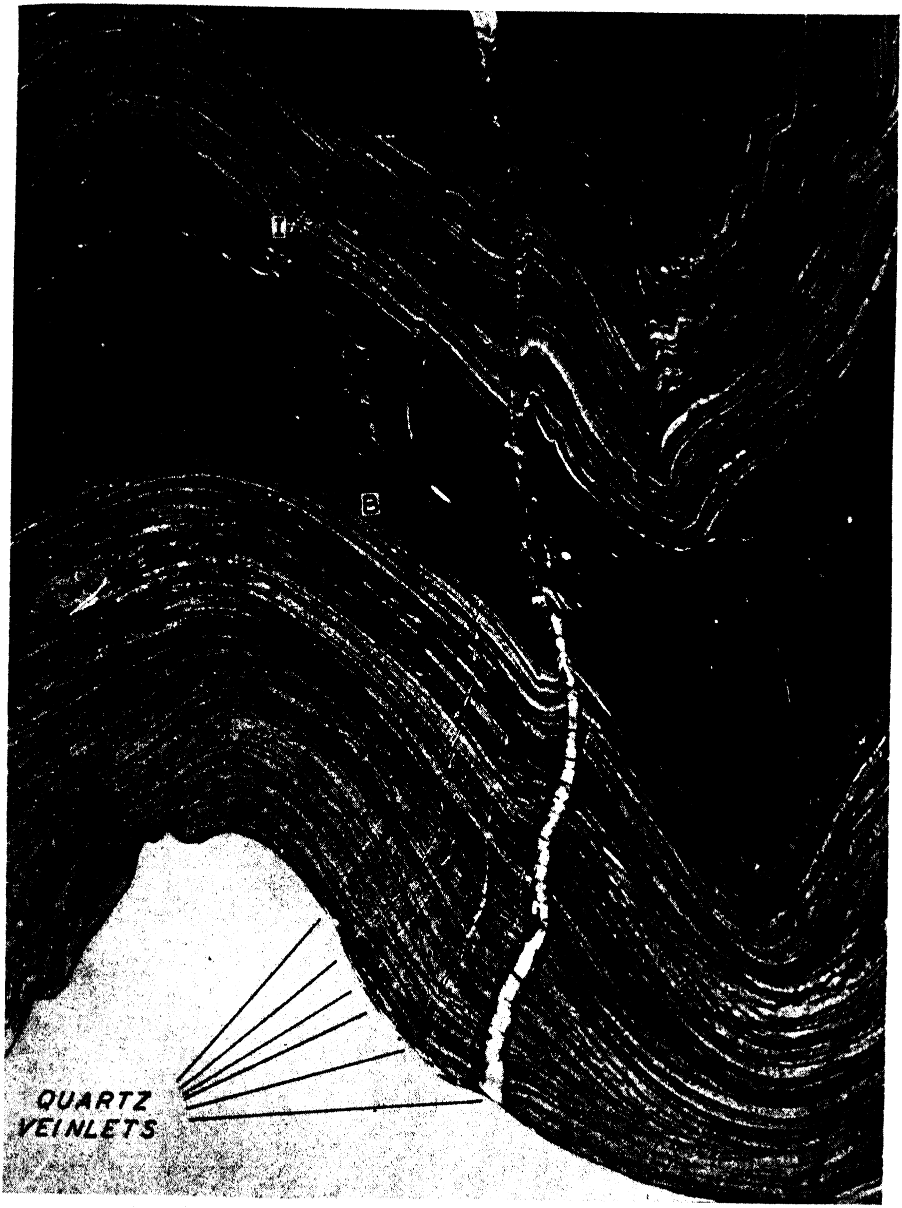


FIG. 1.—Laminated sandstones and slates, Mitta Mitta River, Mitta Mitta, Victoria. Slightly enlarged.

is no evidence of bedding-slip having occurred along any stratification plane within the slate, but the veinlets show strong general curvature indicative of mass flowage of the enclosing argillaceous material.

Under the microscope, it is seen that the quartz has been deformed, first, as a result of differential flowage of the surrounding rock-mass in the general direction of the bedding and, second, by shearing along planes of later development, lying at an angle to the bedding and constituting a type of fracture or slip-strain cleavage (Fig. 4). The microfoliation, as shown by the parallel development of micaceous minerals, is parallel to the bedding in this rock and is quite unrelated to the planes of fracture cleavage.

The amount of flowage since the formation of the veinlets may be inferred from the bending shown by them, and it will be seen that there was a progressively greater amount from about the middle of the right-hand limb of the anticline, toward the axis of the fold. This flowage is explicable according to the usual interpretation of similar folding, in which incompetent rocks undergo mass movements from the limbs toward the axes of folds.

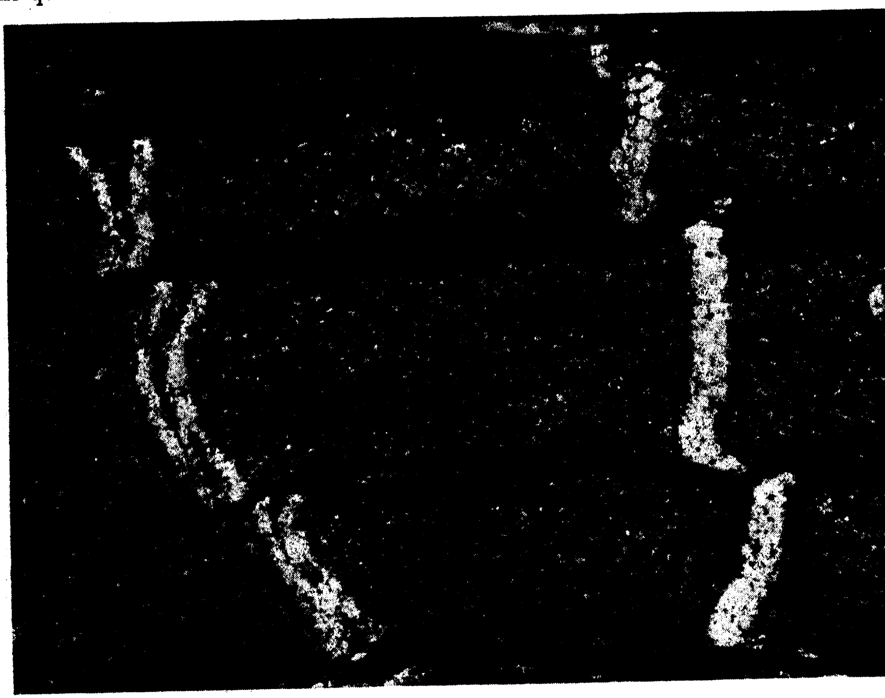


FIG. 2.—Quartz veinlets in laminated sandstone, showing displacements by bedding-slip during folding. ×25.

Slip-folding.—As may readily be seen in Figures 1 and 5, much information about the displacements within the slate can also be obtained from a study of the very thin laminae of slightly different color that are revealed when a smooth surface is flooded with water or oil. By this means, it is possible to reveal struc-

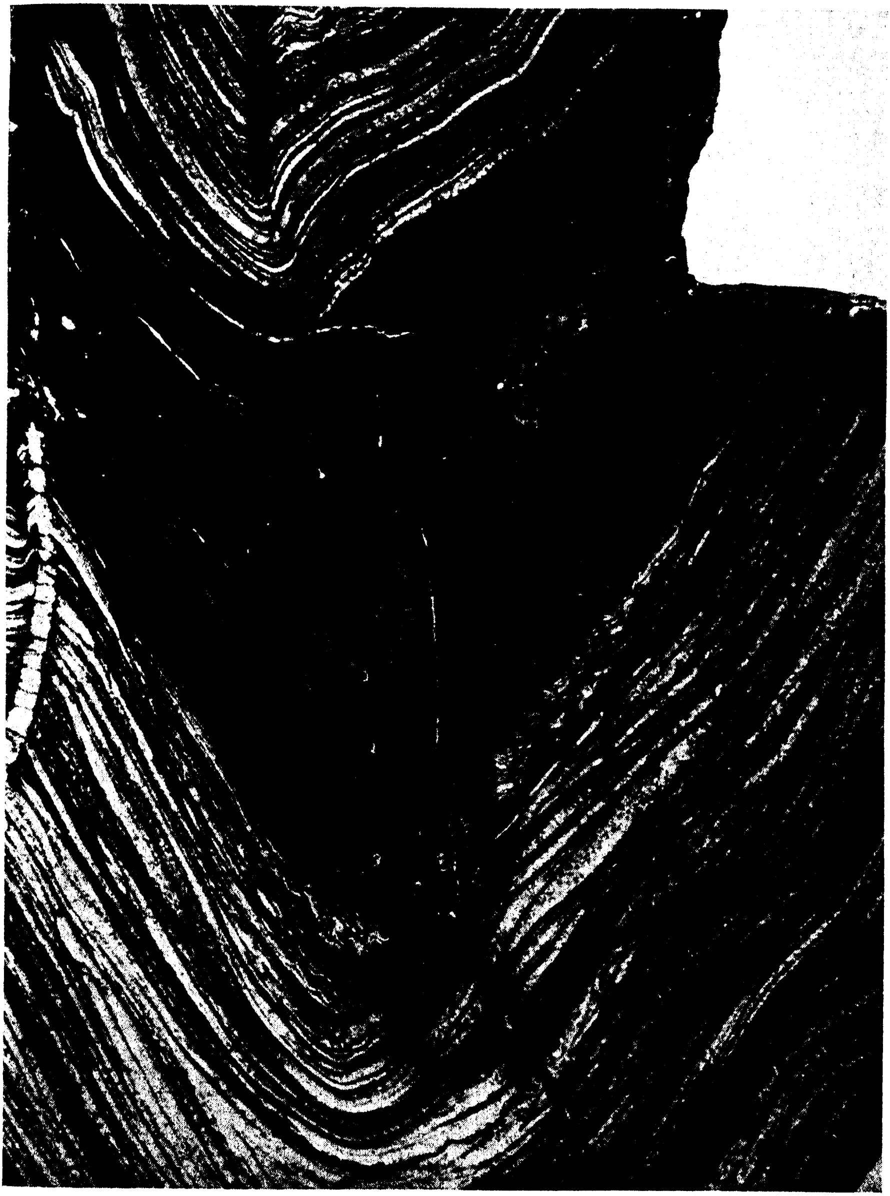


FIG. 3.—Portion of lower right-hand corner of Fig. 1 enlarged to show concentration of slip-planes in the right-hand unit of the syncline (compare Fig. 7.) $\times 2$.

tures that are practically invisible in the untreated specimen.

Evidence for slip-folding is shown, the laminae exhibiting a succession of displacements along closely spaced, statistically parallel shearing planes, whereby a general foldlike form is produced.

FIG. 4.—Quartz veinlet in slate, showing effects of shattering by flowage parallel to the bedding (horizontal); planes of false (strain-slip) cleavage are of later development. ×25.

We have, therefore, evidence for three different types of internal readjustments in this rock, presumably all related to folding. The bedding-slip phenomena in the laminated sandstones indicate that these folded mainly by flexing, accompanied by only minor internal readjustments of grain structure connected with the slight plastic deformation of the laminae above and below their neutral surfaces. Mass flowage of the incompetent argillaceous material occurred during the earlier stages of the formation of the folds, but finally this material yielded by slipping along closely spaced shearing planes, producing slip-folds, while the purer arenaceous material continued to yield by flexing. The relationships where arenaceous and argillaceous laminae alternate and are subequal in importance in the makeup of the rock are, however, not so clear cut. We find, for example, that thin arenaceous laminae embedded in the argillaceous material are sheared through along the cleavage planes, while other arenaceous laminae show various intermediate stages between such clear-cut shearing, shearing with marked "drag," and small-scale flexural folding. Certain generalizations concerning these phenomena are discussed below.

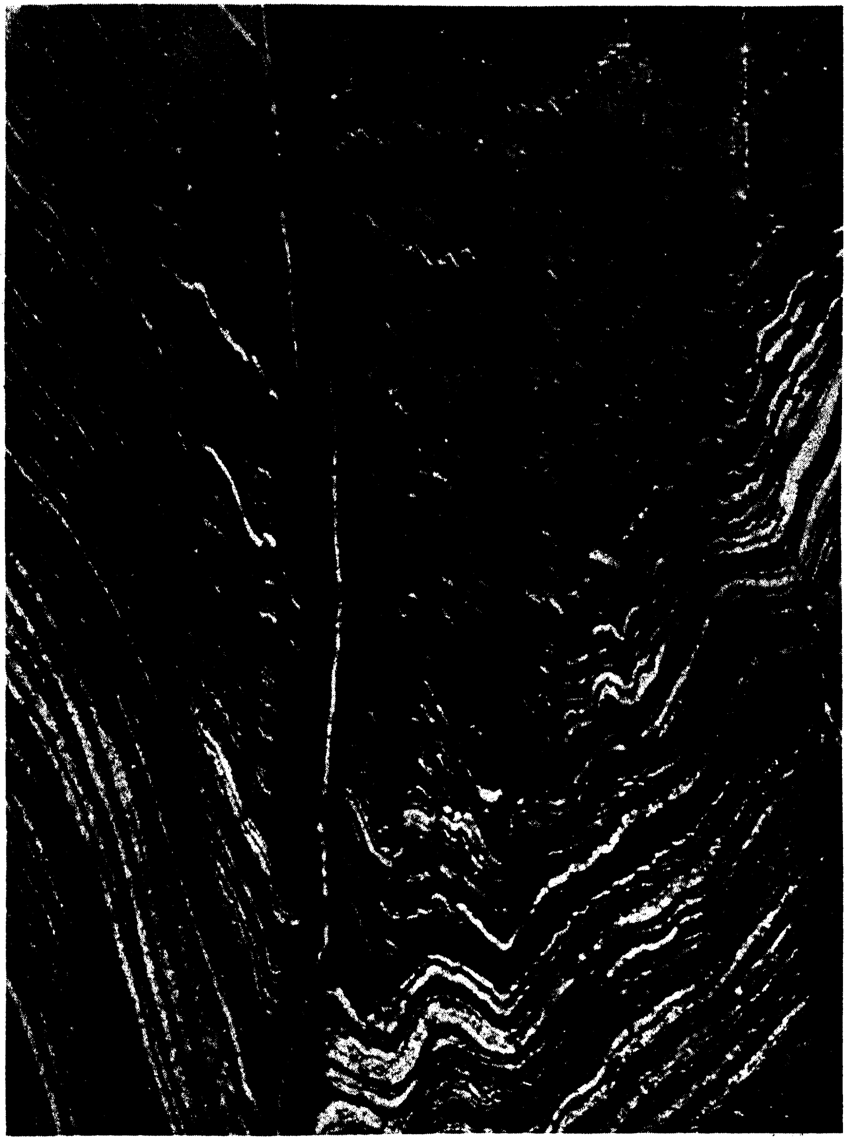


FIG. 5.—Portion of top right-hand corner of Fig. 1 enlarged to show slip-folding in laminated slate. ×4.

Spacing of slip planes.—Study of the arenaceous laminae embedded in the slate and sheared through along the cleavage planes leads to the conclusion (Item 1) that, the thicker the sandy lamina involved, the wider the spacing of the planes of slip (fracture cleavage) within it (see Fig. 6).

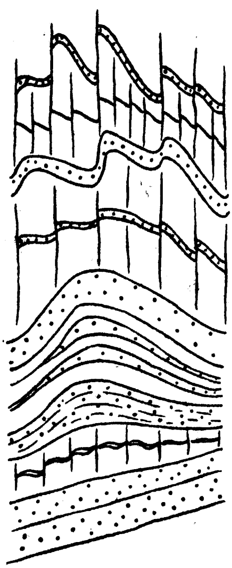


FIG. 6.—Sketch to show relationship of shear planes and folds to lithology in laminated slate and sandstone.

This relationship indicates a connection between the spacing of slip planes and the physical properties of the sheared rock. For a given amount of deformation, if few shearing planes are developed, the movement along each must be greater than if a larger number of planes is active. If, when a few shearing planes have formed, it is easier for slip to continue along these than for new planes to originate, the spacing will remain wide and the movement on each plane will be great. In a sandstone the internal friction is presumably great, so that it should be easier to continue movement along a shearing plane already formed, and on which the friction may have been reduced by attrition of angular grains, than to originate a new plane in such granular rock. It is probable that the thicker sandy laminae in the rock in question contain less argillaceous material than the slightly darker and finer-grained thin sandy laminae, so that the internal friction in the thicker laminae would be greater. In view of the relationships expressed in Item 1 above, it therefore appears that the spacing of shearing planes in the various laminae present is directly related to the internal friction. This relationship, however, may not apply to other examples of plastically deformed rocks. It will be affected by work hardening, especially in the plastic flow of crystalline rocks, and, for a given rock under various tectonic conditions, by the amount of plastic deformation that is required.

Transition from slip to flexural folding.—On this topic the generalization (Item 2) may be made that with an approximately constant amount of slip along any one shearing plane traversing several arenaceous laminae (as shown in Fig. 6):

1. Thin laminae are sharply separated on either side of the plane;
2. Thicker laminae show marked "drag" which tends to provide a shear link between the displaced parts;
3. Thicker laminae still are "folded," with a marked thinning of the sheared limbs of the folds;
4. In the dominantly arenaceous parts the shears die away.

These relationships appear to be connected, first, with the ability of any particular layer or group of laminae to yield by bedding-slip. We note (*a*) that bedding-slip has occurred between the laminae in dominantly arenaceous material

and (*b*) that bedding-slip did not occur in dominantly argillaceous material.

Most probably such slip was inhibited within the latter because the adherence of the rock across the stratification planes was stronger than the shearing-stress component along them. The argillaceous material was, therefore, statistically isotropic in its reaction to stress, and the shearing planes developed in it must, therefore, have a definite geometrical relationship to the stress-axes. In the dominantly arenaceous material, on the other hand, the resistance to shearing is less in the stratification planes than in the rock itself. Such material is anisotropic in its relation to stress; shearing tends to be restricted to the bedding planes (bedding-plane slip) and has no constant geometrical relationship to the stress-axes.

Intermediate stages between slip and flexual folding (Item 2*b*, *c*) occur where the arenaceous and argillaceous laminae are subequal in the composition of the rock.

Consider first the condition with sandy laminae that yield along the shearing planes by flexing, with stretching of the sheared limb of the flexure. In such shear flexures an individual shearing plane in the argillaceous material, on passing into the sandstone, becomes a shear zone with lateral drag effects. Second, with thicker sandstones, the necessary dip shift along a shearing plane or group of planes is achieved by a hinge rotation, resulting in a flexural fold. The formation of such folds involves bedding-slip, which commences to operate when the resistance to shearing in a thick sandy lamina becomes greater than the adhesion of the argillaceous to the arenaceous material. This relationship thus sets a limit to the spacing of shearing planes in the sandy laminae, above discussed, and determines whether slip-folding or flexural folding will operate in any part of the rock mass. Furthermore, where the arenaceous and argillaceous laminae occur in about equal amount, the folding is of the flexural-slip type, both processes being of approximately equal significance. This term is also appl cable to the specimen considered as a whole.

MECHANICS OF FOLDING

In attempting to unravel the rather complex structural history of the specimen, it is necessary to work backward from the final stage, that of slip-folding. It may be observed, in considering the attitude of the stratification planes in the slip-folded argillaceous material, that the very thin laminae in the core of the syncline are parallel with the bedding planes in the left-hand limb of this fold (see Fig. 1 [top right-hand corner] and Fig. 5).

This relationship implies that these laminae, and also the laminated sandstone beds over and underlying them in the right-hand limb of the fold near the axis, were, before the formation of the slip fold, parallel to the beds in the left-hand limb. As the laminae are traced farther to the right, however, they gradually change their attitude until they are parallel to the beds in the rest of the right-hand limb. A similar condition is to be seen in the left-hand limb of the anticlinal fold.

These features are interpreted as follows: At a certain stage in deformation, before slip-folding was initiated, the rock as a whole developed similar flexural folds, involving flowage and bedding-slip. The end of this stage was reached with the anticlinal and synclinal axes in the positions A^1 and S^1 shown in Figure 7. The quartz veinlets were injected before

the completion of this stage of the folding.

Further deformation, resulting in a shifting of the anticlinal and synclinal axes closer together, was achieved by continued flexural folding of the dominantly arenaceous material with slip-folding in the argillaceous material, as shown in the figure. The important point is that, during this stage, the left-hand limb of the syncline remained fixed while the right-hand limb was rotated, and the fold axes closed in. With a slip-fold in

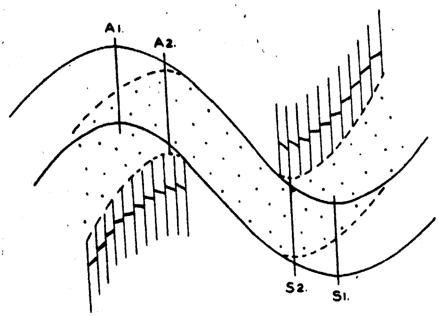


FIG. 7.—Suggested stages in folding. A^1, S^1, anticlinal and synclinal axes before slip-folding; A^2, S^2 positions of axes after slip-folding with a stationary middle limb.

which both limbs rotate about a fixed axis, the throw on the shearing planes is right-hand-up in one limb and left-hand-up in the other.[5] In the example here described, on the other hand, the displacements are right-hand-up along all shearing planes in the vicinity of the fold axes.

CONCLUSION

In their different reactions to stress, including shearing, shear-flexing, and flexural folding, the various laminae in the rock under discussion exhibit different degrees of competency, affording a clear analogue on a minute scale with the structures of larger folds. Obviously, none of the laminae can be termed "competent" in the sense used by Willis,[6] that is, competent to "carry up the weight of overlying strata"; but C. M. Nevin[7] has pointed out that "no bed or formation is sufficiently competent to support itself, let alone lift a considerable burden of overlying rock, under conditions common to folding." The fundamental principle of competency, especially of relative competency, may perhaps best be considered in terms of those properties that determine the geometry of strata in folding. For the rock under discussion, the properties involved would appear to be (*a*) the internal resistance to shear across the bedding; (*b*) ability to yield by slip along bedding planes; and (*c*) ability to yield by mass flowage.

Strata that can yield readily by flowage exert little influence on the general form of folds, except as they transmit hydrostatic pressure, afford lubrication, and, as it were, "fill the spaces" between more competent strata. They accommodate themselves freely to the geometrical forms assumed by the other rocks present and are at the lowest grade of competency.

In strata that yield by slip-folding we have to deal with a more regulated type of plastic flow, in which the displacements, especially along the shearing-planes, bear a definite geometrical relationship to the stress-axes. In the rock under discussion slip-folding and flexural folding are related, but, in the sense of influencing the geometry of the fold, the argillaceous material, as is normally the case, was clearly less competent than the arenaceous at the stage of

[5] B. Sander, *Gefungekunde der Gesteine* (Vienna: Julius Springer, 1930), Fig. 145, p. 269.

[6] *Geologic Structures* (New York: McGraw-Hill Book Co., 1934), p. 79.

[7] *Principles of Structural Geology* (New York: John Wiley & Sons, 1942), p. 50.

slip-folding. As we have seen, mass flowage preceded slip-folding in the argillaceous material, so that presumably some change in the constitution and physical properties of this material took place at a certain stage in the deformation. This stage probably marks the formation of reconstitution minerals in the parts that are now slate. Although there may be slates that do not exbihit slip-folding, as soon as an argillaceous rock does yield in this way, it becomes a slate, and the planes of slip constitute a variety of false cleavage. In the present example the cleavage is an axial-plane cleavage, but, as the above analysis shows, it comprises only one set of shearing planes, which were not predetermined by any inhomogeneity in the rock and, therefore, must be regarded as lying obliquely to the *AB* plane of the strain ellipsoid.

This in turn indicates the necessity for some reservation in accepting the widely held view that "the axial planes of folds are essentially perpendicular to the least strain axis."[8]

[8] M. P. Billings, *Structural Geology* (New York: Prentice-Hall, Inc., 1942), p. 216.

AN OCCURRENCE OF "CAVE PEARLS" IN A MINE IN IDAHO[1]

J. HOOVER MACKIN
U.S. Geological Survey

HOWARD A. COOMBS
University of Washington, Seattle

ABSTRACT

Pisolites up to 13 mm. in diameter occur in depressions in a veneer of calcium carbonate which mantles rubble in an abandoned mine. They were formed by precipitation from mine water, and the high polish that characterizes some specimens is due to agitation caused by dripping water. The period of formation is 35–42 years.

INTRODUCTION

Pisolites, formed by the accretion of concentric layers of calcium carbonate around a nucleus, have been described by Frank L. Hess[2] from specimens collected by Lee in the Carlsbad Caverns, by W. D. Keller[3] from a cave in Missouri, and by Norbert Casteret[4] from a cave in France. Edvard Erdman[5] has discussed their formation in a mine in Sweden, and W. H. Twenhofel[6] has called attention to the growth of somewhat analogous oölites in a water-heating system. The pisolites treated here differ in some respects from those described in these earlier papers, shed light on the origin of the high polish that characterizes the type of pisolites known as "cave pearls," and are of special interest because the period required for their formation is known.

The pisolites were found by Mackin in the course of an examination of an abandoned prospect adit in the Iron Mountain district in southwestern Idaho—the lower adit on the south side of the West Fork of Crawford Gulch (S.W. ¼, sec. 12, T. 14 N., R. 6 W.). The petrographic studies, photomicrographs, and pictures were made by Coombs.

According to Messrs. John Siegwein and Frank Mortimer, of Weiser, Idaho, the adit in question was opened in 1901. It is possible that it may have been cleaned out in the interval between 1901 and 1908, when patent was issued on the mining claim. So far as is known, no work has been done on the property since that time. In other words, the period of development of the pisolites is 35–42 years (1901 or 1908 to 1943).

OCCURRENCE

The country rock in the Crawford Gulch prospect is a coarse-textured marble which is locally mineralized at

[1] Published with permission of the director, Geological Survey, U.S. Department of the Interior.

[2] "Oölites or Cave Pearls in the Carlsbad Caverns," *No. 2813 of Proc. U.S. Nat. Mus.*, Vol. LXXVI, art. 16 (1930), pp. 1–5. For an early mention of the Carlsbad oölites, and an illuminating discussion of both organic and inorganic coated stones, see Heinrich Schade, "Zur Entstehung der Harnsteine und ähnlicher konzentrisch geschichteter Steine organischen und anorganischen Ursprungs," *Zeitschr. f. Chem. u. Ind. d. Kolloid.*, Vol. IV (1909), pp. 175–80, 261–66. This important article, apparently little known to geologists, was called to the writer's attention by W. H. Bradley. A digest was prepared by Gordon L. Bell.

[3] "Cave Pearls in a Cave near Columbia, Missouri," *Jour. Sed. Pet.*, Vol. VII (1937), pp. 263–65.

[4] *Ten Years under the Earth* (New York: Greystone Press, 1939), pp. 204–6.

[5] "Stalagmit och pisolitartade bildningar i Hoganas stenokolsgrufa, Shone," *Geol. fören i Stockholm förhandl.*, Vol. XXIV (1902), pp. 501–7.

[6] "Oölites of Artificial Origin," *Jour. Geol.*, Vol. XXXVI (1928), pp. 564–68.

and near the margins of a quartz-diorite intrusive body; the adit was driven in search for a sulphide vein. The portal is partly caved; and, though the air is

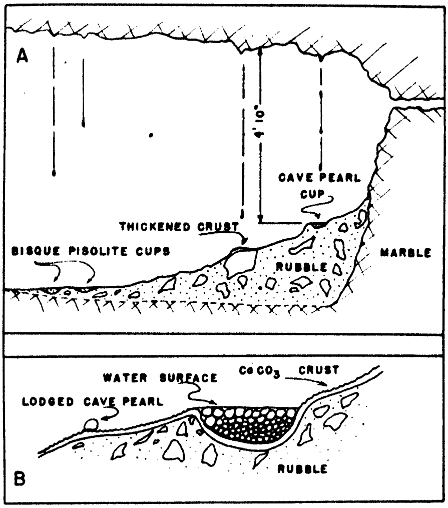


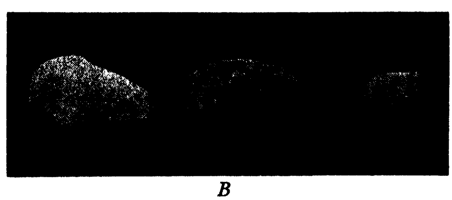
FIG. 1.—*A*, diagrammatic section of adit, to show occurrence of "cave pearls." *B*, detail of "cave-pearl" cup.

fresh, there is no free circulation. The walls and floor of the adit in a segment 80–120 feet from the portal are more or less completely covered by a rough calcium-carbonate crust ranging from a thin film to several centimeters in thickness; the surface of the crust is kept wet by seepage from the walls and dripping from the roof. The pisolites occur in shallow, cuplike depressions in rubble on the floor and at the base of the walls (see Fig. 1).

Two of the cups, 5–7 cm. in diameter and 3–5 cm. in depth, contain pisolites whose brilliant polish amply justifies the term "cave pearls." In both cups ten to fifteen round to subround, relatively large "pearls" (10–13 mm. in diameter) form a distinct surface layer flush with the rim (see Fig. 2, *A*, right). The underlying pellets are smaller (2–4 mm.) and have corners and edges more highly polished than the flat or re-entrant faces (see Fig. 2, *A*, center). The interior walls of the cups are smoothly undulating surfaces exhibiting the same high polish as the contained pellets, contrasting markedly in this respect with the surrounding rough-surfaced crust.

Both of the "cave-pearl" cups are on a slope above the general level of the floor of the adit (Fig. 1, *A*); about ten of the larger-sized pellets, evidently derived from the cups, are lodged on the descending slope and on adjoining parts of the floor. All are cemented in place, some still retaining a brilliant polish on the upper parts, some having the polish

A



B

FIG. 2.—*A*, the small "cave pearls" in the center are from 2 to 4 mm. in diameter; the larger specimens on the right vary from 10 to 13 mm. in diameter. The bisque-surfaced pisolite on the left shows polished interior growth layers. *B*, rough and bisque-surfaced pisolites. The specimen in the center is 25 mm. in diameter.

frosted over in greater or less degree, and some being so completely crusted by calcium carbonate as to appear as rough projections above the adjacent surface.

Of nine other nests of pisolites observed in the adit, four contain subrounded pellets showing some polish on nodes and edges but with a general surface texture similar to that of unglazed porcelain or bisque china. The remaining cups are shallow and less regular in form and are set apart from the others by the fact that some of the contained pisolites are cemented to the floor and walls; most of the free pisolites are rough surfaced, some are smooth, but none show areas of polished surface. Finally, various irregular shallow depressions on the floor of the adit contain rough, subangular pisolites, the majority cemented in place.

The pisolites may be grouped, then, into three main types: (1) the highly polished "cave pearls" (Fig. 2, *A*, center and right); (2) bisque-surfaced pisolites, including those locally polished on nodes and edges (Fig. 2, *B*, center and right); and (3) rough-surfaced pisolites (Fig. 2, *B*, left). There are, of course, gradations between these types. Some bisque-surfaced pisolites show inner-growth rings which are highly polished (Fig. 2, *A*, left).

PETROGRAPHIC DESCRIPTION

All of the pisolites, whether they be smooth or rough, have a nucleus of some type; but these vary greatly in size and shape. Some of the nuclei are covered by a very thin coating of calcium carbonate; in others the coatings are extremely thick and represent by far the greater volume of the "pearl." The smooth and more rounded types usually have a thicker coating than the rough subangular or noded types (see Fig. 3).

In the twenty-five specimens sectioned for petrographic study, no two have nuclei of exactly the same composition. The most common type of nucleus is a fragment of garnet-calcite-magnetite rock; other types are composed of various lime-silicate minerals or of ferruginous marble with minor amounts of pyrite and epidote. A yellowish garnet of the grossularite variety is the most abundant mineral. One specimen is a single-zoned garnet crystal enclosed by a comparatively thick covering of calcium carbonate (see Fig. 3, *A* and *B*).

Wrapped concentrically around the nuclei are rims of calcium carbonate ranging from a fraction of a millimeter to 5 mm. in thickness. Some of the sections and sliced pisolites were stained with Meigen's solution, known pieces of calcite and aragonite being used as controls. Many of the very thin layers proved to be aragonite, whereas the coarser and thicker layers were usually composed of calcite. The fine dark bands in the central portion of Figure 4, *A*, are stained aragonite. Figure 4, *A*, is the only photomicrograph showing a stained section; in the other photomicrographs (*B*, *C*, and *D*) the darker bands are due to impurities in the calcite.

The number of layers in a single pisolite ranges from a few to thirty or more (see Figs. 3 and 4). The thick coating on the specimen shown in Figure 4, *A*, contains approximately thirty bands, as counted under the microscope, although that number may be difficult to distinguish in the photomicrograph. The thin coatings on the specimens in Figure 3, *C* and *D*, show ten recognizable layers. It will be noted in Figure 3, *C* and *D*, that the thick basal layer of calcium carbonate is not entirely continuous over the garnet-epidote-calcite nucleus. The outer layer of the specimen shown in Figure 4, *C*, is much thicker than the earlier-formed layers and is made up of calcite crystals in semiradiating masses. A tendency toward a rounding of outlines

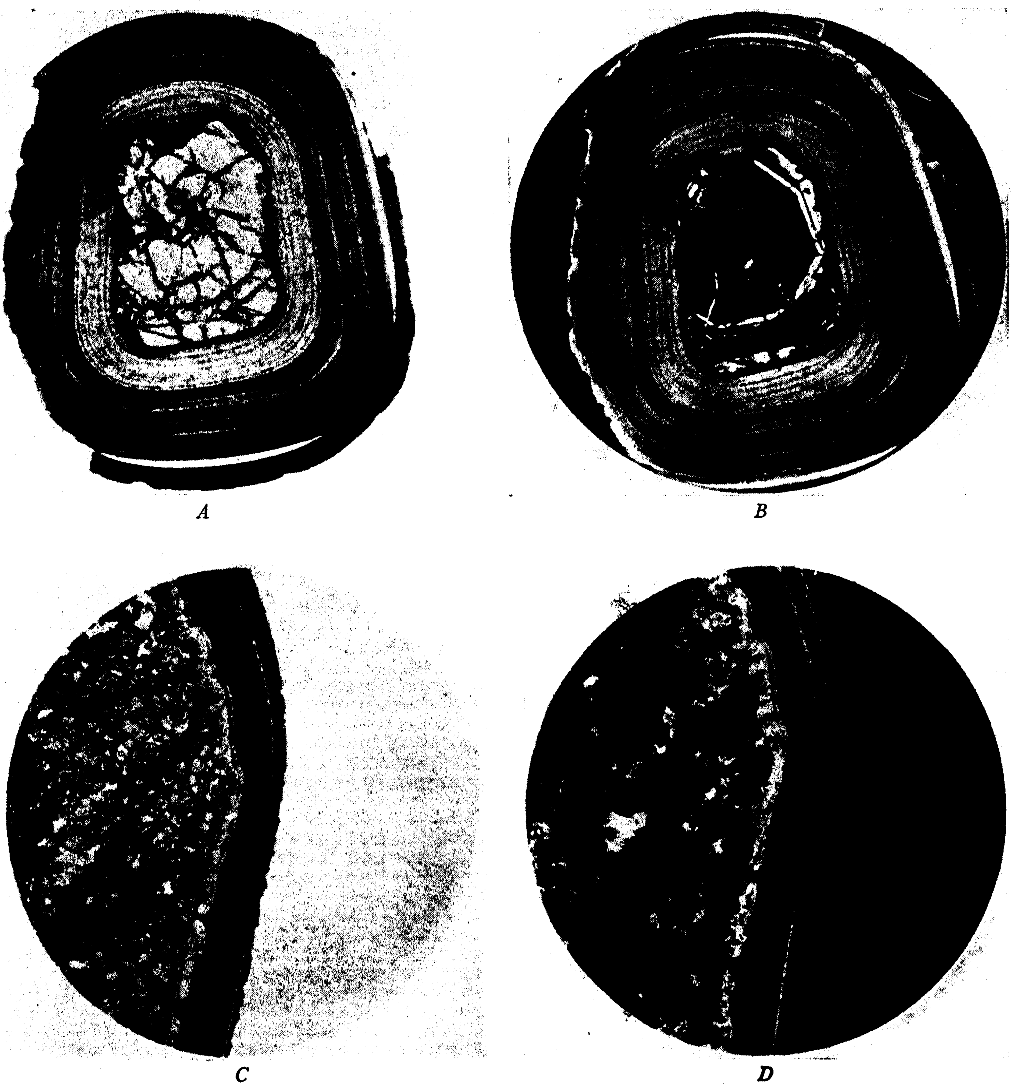


FIG. 3.—*A*, "cave pearl" with zoned garnet nucleus, plane light (greatest diameter, 4 mm.). *B*, same as *A* but crossed Nicols. *C*, bisque-surfaced pisolite; the nucleus is composed of garnet, calcite and epidote; the rim is 1 mm. thick. Note that the first layer of calcite is not continuous over the nucleus. *D*, same as *C*, but under crossed Nicols.

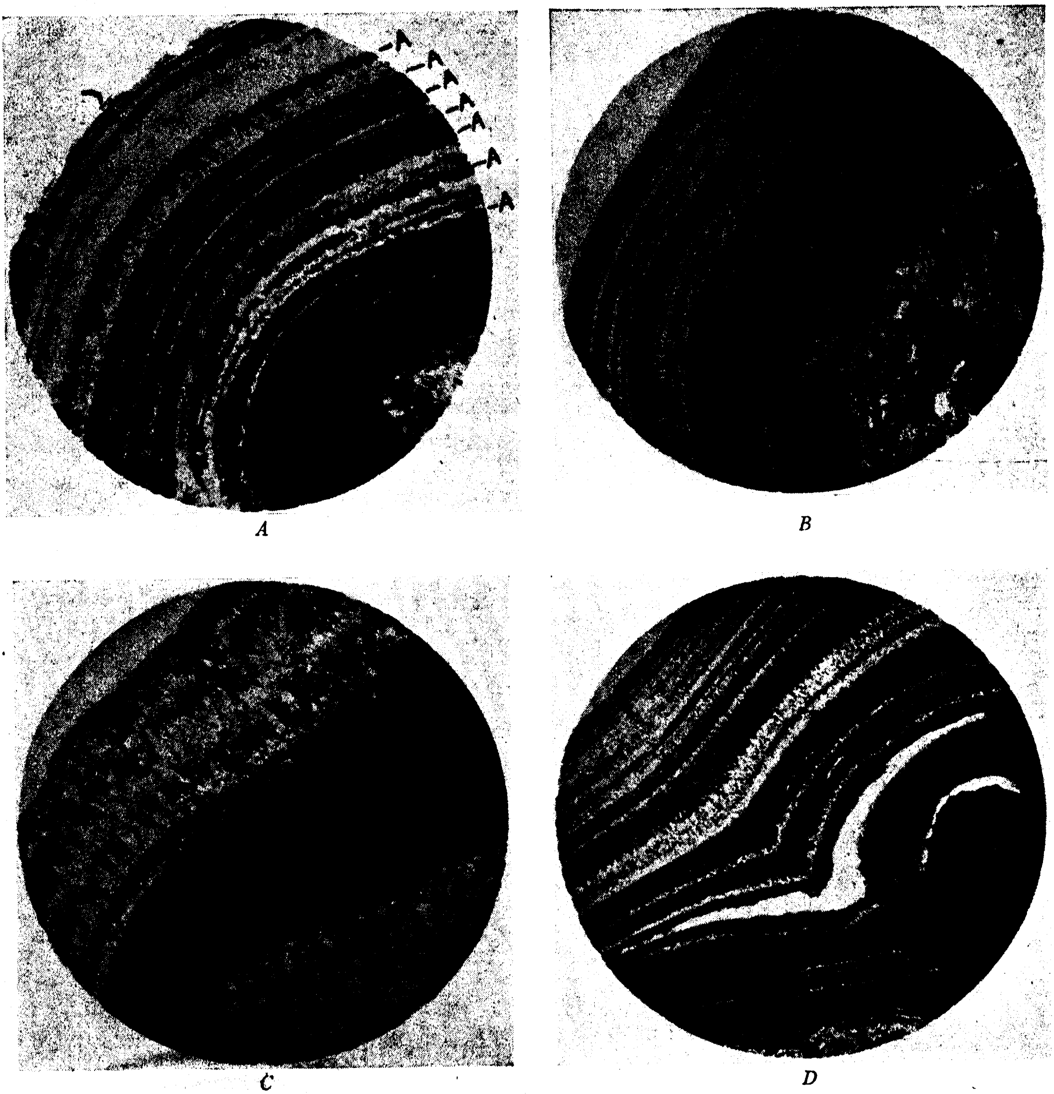


FIG. 4.—*A*, bisque pisolite; stained to show aragonite bands (marked *A*). Width of section seen in photomicrograph, 1 mm. *B*, bisque pisolite; the darker bands are due to impurities in the calcite. Approximately thirty rings can be counted on this specimen. Width of section in photomicrograph, 1 mm. *C*, bisque pisolite—a very fine early series of layers covered by a very coarse outer layer of calcite. Width of section, 1 mm. *D*, smooth pisolite; the large node has been smoothed over by variations in thickness of later layers. Width of section, 1 mm.

with growth of the calcium-carbonate coatings is shown in a number of the photomicrographs.

In Figure 4, *C*, an angular corner of the nucleus was evidently changed to a smooth curve by the first few growth layers. Figure 4, *D*, shows the manner in which a node and the associated re-entrant angle formed during one stage of growth have been smoothed out by variation in thickness of the layers formed during later stages. The clear white band of calcite near the outer margin of the coating in Figure 3, *A* and *B*, is thick on the relatively flat sides of the specimen and thin or missing altogether on the corners.

HYPOTHESIS OF ORIGIN

After the tunneling operations were halted, the depressions were formed in loose rubble on the floor and along the walls of the adit by erosive action of localized drippings of water from the roof. It is evident that the development of some of the cups involved merely a smoothing of initial irregular depressions in the mine waste. Where the rubble consisted of fragments too large for erosion by impact and splash action, no depressions were formed (Fig. 1).

The floors and walls of the cups and adjoining surfaces of rubble and rock kept wet by mine waters were veneered by calcium carbonate; rims of some of the cups were raised by precipitation from overflow waters. Rock fragments and grains in the cups were similarly coated by calcium carbonate and, if agitation by dripping from the roof was sufficient to keep them in motion, became free-coated stones, or pisolites. The degree of smoothness or polish on the surfaces was dependent on whether the particles were merely rocked to and fro or whether they were turned over and over. This, in turn, depends upon whether the original particle was tabular or more or less equidimensional in shape, on the vigor of agitation, on the shapes of companion particles in the same cup, on the shape of the cup, and other circumstances. If accretional growth was not accompanied by motion of the particle for any considerable period, the fragments (or pisolites developed earlier) were cemented to the bottoms and floors of the cups.

This hypothesis provides an explanation for many of the features of the pisolites. The gradational sequence from a few cups with highly polished "cave pearls," through a greater number of cups with smoothed but only locally polished pisolites, to a still greater number of depressions with rough, free pisolites and many fixed pisolites is believed to be due to variation, from cup to cup, in the several controls mentioned above. The rate of dripping and, especially, variations in the rate are particularly important.

At the time of the examination (August, 1943), after at least six weeks without rain, both of the cups containing "cave pearls" were receiving drops from the roof at intervals of 5–15 seconds. To aid in determining whether the "pearls" were in motion, several were removed, marked with pencil dots, and carefully replaced in their original position; observation over a few minutes' time showed only a very slight movement of one of the marked specimens. To duplicate conditions of more rapid dripping, which certainly occur during the wetter seasons of the year, other waters from the roof were diverted so as to increase the rate to intervals of approximately 1 second; at that rate the water in the cups was in constant agitation, and several of the "pearls" were placed in motion

(or their motion was increased) so that they rotated about one-quarter turn in about 2 minutes.

It is believed that these slow rotational movements, which involve a rubbing-down or buffing of each film of calcium carbonate as it is deposited, are responsible for the high polish that characterizes the "pearls." The tendency for the "pearls" to assume a rounded shape is evidently due, in large part, to the relatively greater rubbing-down of the corners and edges than of the flattish or re-entrant faces (see Figs. 3, *A* and *B*, and 4, *C* and *D*).

The distinct layering in the coatings is probably due to variation in agitation and nourishment (in both composition and supply), resulting from variation in the rate of dripping. While it is not likely that the control is strictly seasonal, it is interesting to note that the thirty-odd layers in some of the larger cave-pearl specimens compare closely with the known period of growth of the "cave pearls" (1901 or 1908 to 1943, 35–42 years).

In view of the fact that the cups were essentially brimful with "cave pearls" at the time of the examination, it appears that the "pearls" lodged on the slopes below the cups were dislodged from them as a result of slow accretional growth of all of the contained pellets.

One of the cups containing bisque-surfaced pisolites was receiving drips at approximately 10-second intervals; but in the majority of cups containing pisolites of this type the interval was much longer, and some received no drips during a brief period of observation. An intermittent rocking motion was noted in one cup holding bisque-surfaced pisolites. The fact that the corners and edges of some of the bisque pisolites are polished testifies to mutual rubbing, probably not accompanied by complete rotation. It is likely that with continued growth and rounding of outlines some of the bisque-surfaced forms may change into "cave pearls."

Some of the poorly defined depressions on the floor of the adit were receiving drips at varying rates, but many were supplied with water (during the dry season) only from seepage; most of the pisolites in depressions of this type are rough surfaced, and many are cemented in place. The inference is that continued nourishment by seepage waters during periods of no dripping and agitation tends to cause fixation.

There is evidence of marked changes in the rate and type of accretional growth in several of the cups. Figure 4, *C*, shows a change from early, finely banded growth layers (probably yielding a polished or bisque pisolite) to a thick outside layer of radial calcite yielding a rough exterior surface. Specimen 4, *D*, shows early uniform growth rings. Later, probably during a period of rest with lowered water level, the pisolite developed a node; and still later the growth habit changed again in such manner as to produce a smooth external form. As indicated earlier, the bisque-surfaced pisolite on the left in Figure 1, *A*, shows high polish on earlier growth rings. Changes of this type are believed to be due to variations in nourishment and agitation resulting from shifting in the positions of local orifices through which ground water enters the adit.

CONCLUSIONS

The brilliant polish that characterizes the Idaho "cave pearls" is believed to be the result of a buffing or rubbing-down of accretional layers of calcium carbonate as they are deposited, the buffing action being due to slow rotational movement

induced by agitation caused by dripping ground water. Since a rounded form is an essential prerequisite for such slow rotational movements, it is likely that the cave pearls described here did not have polished surfaces during very early stages of calcium-carbonate growth around nuclei consisting of angular fragments of mine waste. In other words, rounding of the pisolites by accretional growth and selective wear at edges and corners probably preceded, and made possible, the attainment of the high polish. This does not mean that all of the bisque- or rough-surfaced pisolites in the adit will eventually become polished, for the tendency to become smoothly rounded with growth is not systematic but depends upon a number of factors which vary from cup to cup—many once-free pisolites are now firmly cemented in place.

The pisolites described here were formed during a period not exceeding 42 years, the time interval since the adit was driven.

The theory of origin of the cave pearls described here corresponds in general with that proposed by Casteret for "oölites or cave pearls" occurring in natural caverns in France. Casteret states that "the fragments .... roll perpetually in the whirlpool [as they are being coated] .... and become almost perfectly spherical." Lee and Hess mention observing only a "slow rocking motion" in the cave-pearl nests in the Carlsbad Caverns. The rocking motion will suffice to prevent fixation, but it is not clear how it will produce the brilliant over-all polish that distinguishes "cave pearls" from the other calcium-carbonate precipitates formed by inorganic processes.

A GEOMAGNETIC SURVEY OF SOME BLADEN COUNTY, NORTH CAROLINA, "CAROLINA BAYS"

JOHN C. McCAMPBELL
Rutgers University, New Brunswick, New Jersey

During a recent field season a magnetometer survey comprising more than five hundred stations was carried out over approximately 15 square miles of previously unsurveyed area of the coastal plain of North Carolina. The area lies in Bladen County, about 9 miles northwest of Elizabethtown. Included within it are two single and one double oval or elliptically shaped depressions commonly referred to as "Carolina Bays."

One of the several hypotheses for the formation of the "Carolina Bays" is that proposed by F. A. Melton and William Schriever,[1] who were of the opinion that the bays resulted from the impact of a great many meteorites. A test of this hypothesis is the determination of the presence or absence of appreciable amounts of magnetic materials buried immediately below or close to the bays. Such materials should produce magnetic disturbances or "highs" that would be measurable at the surface by magnetometric surveying.

[1] "The Carolina 'Bays'—Are They Meteorite Scars?" *Jour. Geol.*, Vol. XLI (1933), pp. 52–66.

The plotted results of this survey show that pronounced magnetic highs are present. As shown in Figure 1, they are distributed as follows: high *A* seems associated with a bay which lies immediately to the west of the area surveyed; high *B* is seemingly associated with the double White Oak Bay; while highs *C* and *D* appear associated with the unnamed bay and Sessom's Bay, respectively.

The observed highs are variable in size, "relief," and position. Each one occurs well outside of the bay proper. Highs *A* and *B* occur to the east and south of the extreme southern end of the bays. High *C* is almost due south of the center of the unnamed bay, while high *D* occurs slightly west of a point immediately south of the southern end of Sessom's Bay.

Although previous reports of similar surveys have relied upon such magnetic data as presented here as a conclusive proof of the meteoritic hypothesis of origin of the bays, the question is here left open and the reader is free to arrive at any conclusion as to their origin that may be warranted by the facts presented.

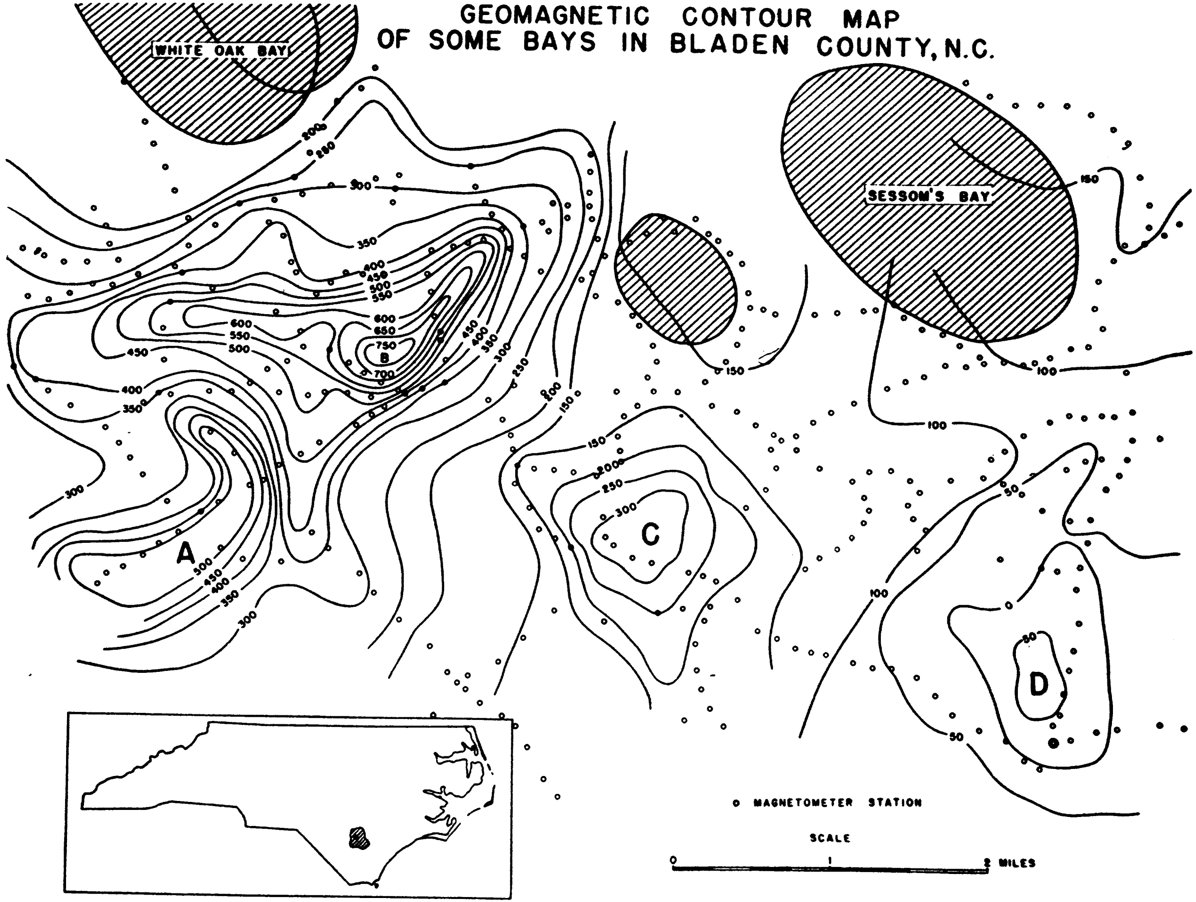


FIG. 1.—Geomagnetic map of a part of Bladen County, North Carolina. Contour interval = 50 gammas

REVIEWS

The Mineral Resources of Africa. By A. Williams Postel. Philadelphia: University of Pennsylvania Press, 1943. Pp. 105; figs. 19. $1.50.

This constitutes No. 2 in a series of African handbooks issued by the Committee on African Studies of the University of Pennsylvania. It is a summary and not an exhaustive treatise. The volume begins with a brief review, mainly statistical, of the world mineral situation, in order that Africa may be viewed in proper perspective. A comparison of African with world production (for 1938) shows that the largest items of Africa's mineral production are as follows:

	Percentage of World's Production
Diamonds	99.0
Vanadium	47.5
Gold	40.0
Phosphate	36.6
Chromium (ore)	31.6
Fluorspar	22.6
Manganese (ore)	21.6
Copper	18.0
Asbestos	16.0
Graphite (approx.)	15.0
Tin	10.8

All others are less than 10 per cent each. Africa has several mineral provinces that are productive. Along the northern rim of the continent, extending from Morocco through Algeria to Tunisia, is the great phosphate-rock belt. Associated with this belt are iron deposits of considerable importance. The largest copper province in the world extends from Katanga, Belgian Congo, southeast into Northern Rhodesia for a distance of 200 miles. This belt is also the largest world producer of radium and cobalt ores. The Bushveld igneous complex in Transvaal, an area of 37,000 square miles, and the Great Dike that runs through Transvaal into Southern Rhodesia, a distance of 330 miles, form another important province. The mineralization in this area includes platinum, chromite, and asbestos. The African diamond province has no equal in the world.

Most of the book is devoted to brief listing of the principal mineral occurrences of economic importance, classed under the usual captions of metals and nonmetallics. Geologic relations are not described, though they are illustrated in some of the figures.

In a brief chapter devoted to water power and supply the significant statement is made that "potential water power may be regarded as Africa's greatest resource." At present, however, not one-hundredth of this vast reservoir of energy is being utilized. In a continent almost devoid of mineral fuels the future of water-power development becomes of overwhelming industrial importance. The total water-power resources of Africa are estimated at three times those of Europe. More than half of this potential power is along the Congo River, mainly in Belgian and French territory.

An appendix deals statistically with the geographical distribution of Africa's mineral production and reserves. A short selected bibliography closes the volume.

This book is useful in portraying the broad features of the African mineral scene and furnishes a good starting-point for those wishing to pursue the subject more deeply.

E. S. Bastin

PUBLICATIONS RECEIVED

Administration Report of the Government Mineralogist for 1942. Colombo: Ceylon Government Press, 1943.

Aerial Photographs and Their Applications. By H. T. U. Smith. New York: D. Appleton-Century Co., 1943.

Analyses of Pennsylvania Anthracitic Coals. By Various Authors. U.S. Bureau of Mines Technical Paper 659. Washington, D.C.: Government Printing Office, 1944.

Analyses of Virginia Coals. By Various Authors. U.S. Bureau of Mines Technical Paper 656. Washington, D.C.: Government Printing Office, 1944.

An Annotated Synopsis of Paleozoic Fossil Spores and the Definition of Generic Groups. By J. M. Schopf, L. R. Wilson, and Ray Bentall. Report of Investigations No. 91. Urbana: Illinois State Geological Survey, 1944.

Antimony in Oregon. By Norman S. Wagner. G.M.I. Short Paper No. 13. Portland, Ore.: Department of Geology and Mineral Industries, 1944.

Artesian Water in Southeastern Georgia. By M. A. Warren. Geological Survey Bulletin No. 49. Atlanta, 1944.

Atoms, Rocks, and Galaxies: A Survey in Physical Science. By John Stuart Allen *et al.* New York: Harper & Bros., 1942.

An Avifauna from the Lower Miocene of South Dakota. By Alden H. Miller. University of California Publications, Bulletin of Department of Geological Sciences, Vol. XXVII, No. 4. Berkeley: University of California Press, 1944.

Bed-Load Transportation in Mountain Creek. By Hans Albert Einstein. U.S. Department of Agriculture, Soil Conservation Service, 1944. Prepared at Greenville Sediment Load Laboratory, Sedimentation Section Office of Research in co-operation with South Carolina Experiment Station, Clemson, South Carolina.

The Blewett Iron Deposit, Chelan County, Washington (with Preliminary Tonnage Estimates). By W. A. Broughton. Report of Investigations No. 10. Olympia: State of Washington Department of Conservation and Development, 1943.

The Blue Asbestos Bearing Banded Iron Formations of the Hamersley Ranges, Western Australia. By Keith R. Miles. Also, The Blue Asbestos Deposits of the Hamersley Range and Their Economic Importance. By J. S. Foxall. Geological Survey of Western Australia Bulletin No. 100. Perth, 1942.

The Buckhorn Iron Deposits of Okanogan County, Washington: Results of a Magnetic Survey. By W. A. Broughton. Report of Investigations No. 8. Olympia: State of Washington Department of Conservation and Development, 1943.

Buteux Area, Abitibi County and Abitibi Territory. By. B. C. Freeman. Geological Report 15, Division of Geological Surveys. Quebec, 1943.

Calumet Island Area, Pontiac County. By F. FitzOsborne. Geological Report 18, Division of Mineral Deposits. Quebec, 1944.

Character and Tonnage of the Turk Magnesite Deposit. By W. A. G. Bennett. Report of Investigations No. 7. Olympia: State of Washington Department of Conservation and Development, 1943.

Chloride in Ground Water in the Coastal Plain of Virginia. By D. J. Cederstrom. Virginia Geological Survey Bulletin 58. University, Va., 1943.

Coal Resources of the Kansas City Group, Thayer Bed, in Eastern Kansas. By Walter H. Schoewe. Bulletin 52, State Geological Survey of Kansas. Lawrence, 1944.

The Contribution of Holland to the Sciences. By A. J. Barnouw and B. Landheer (eds.). Querido, N.Y.: American Book–Stratford Press, Inc., 1943.

Corals from the Chouteau and Related Formations of the Mississippi Valley Region. By William H. Easton. Report of Investigations No. 97. Urbana: Illinois State Geological Survey, 1944.

Development and Use of Certain Flotation Reagents. By R. S. Dean and P. M. Ambrose. U.S. Bureau of Mines Bulletin 449. Washington, D.C.: Government Printing Office, 1944.

Developments in Eastern Interior Basin in 1943. By Alfred H. Bell. Press Bulletin Series No.

49, Illinois State Geological Survey. Urbana, 1944.

Differential Thermal Analysis of Clays and Shales: A Control and Prospecting Method. By R. E. Grim and R. A. Rowland. Report of Investigations No. 96. Urbana: Illinois State Geological Survey, 1944.

Directory of Georgia Mineral Producers, 1943. Department of Mines, Mining, and Geology. Atlanta, 1943.

Dolomite Resources of Washington, Part I: Preliminary Report on Okanogan, Lincoln, and Stevens Counties. By W. A. G. Bennett. Report of Investigations No. 13. Olympia: State of Washington Department of Conservation and Development, 1944.

Domestic Coke from Illinois Coals: An Experimental Coke-heated Oven. By Frank H. Reed and Harold W. Jackman. Report of Investigations No. 99. Urbana: Illinois State Geological Survey, 1944.

Economic Aspects of the Blewett–Cle Elum Iron Ore Zone, Chelan and Kittitas Counties, Washington. By W. A. Broughton. Report of Investigations No. 12. Olympia: State of Washington Department of Conservation and Development, 1944.

An Empirical Method of Interpretation of Earth-Resistivity Measurements. By R. Woodward Moore. American Institute of Mining and Metallurgical Engineers Technical Publication No. 1743. Washington, D.C., 1944.

Exploration for Oil and Gas in Western Kansas during 1943. By Walter A. Ver Wiebe. Bulletin 54, State Geological Survey of Kansas. Lawrence, 1944.

Geology and Ground-Water Resources of the Camp Shelby Area. By Glen Francis Brown. Mississippi State Geological Survey Bulletin 58. University, Miss., 1944.

The Geology and Ground Water Resources of the Emerado Quadrangle. By Wilson M. Laird. North Dakota Geological Survey Bulletin 17. Grand Forks, 1944.

Geology and Ground-Water Resources of Finney and Gray Counties, Kansas. By Bruce F. Latta. Bulletin 55, State Geological Survey of Kansas. Lawrence, 1944.

Geology and Ground-Water Supply at Camp McCain. By Glen Francis Brown and Robert Wynn Adams. Mississippi State Geological Survey Bulletin 55. University, Miss., 1943.

Geology and Ground-Water Supply at Camp Van Dorn. By Glen Francis Brown and William Franklin Guyton. Mississippi State Geological Survey Bulletin 56. University, Miss., 1943.

Geology and Ore Deposits of the Superior Mining Area, Arizona, and Supplement. By M. N. Short *et al.* Arizona Bureau of Mines, Geological Series No. 16, Bulletin No. 151. Tucson: University of Arizona, 1943.

The Geology and Underground Waters of Southern Minnesota. By George A. Thiel. Minnesota Geological Survey Bulletin 31. Minneapolis: University of Minnesota Press, 1944.

Ground-Water Conditions in the Neosho River Valley in the Vicinity of Parsons, Kansas. By Charles C. Williams. Bulletin 52, State Geological Survey of Kansas. Lawrence, 1944.

Hydrocarbons of the Uinta Basin of Utah and Colorado. By Clark F. Barb and James O. Ball. Colorado School of Mines Quarterly, Vol. XXXIX, No. 1. Golden, 1944.

Illinois Mineral Industry in 1942. Walter H. Voskuil and Douglas F. Stevens. Report of Investigations No. 94, Illinois State Geological Survey. Urbana, 1944.

Inventory of Mineral Properties in Chelan County, Washington. By Marshall T. Huntting. Report of Investigations No. 9. Olympia: State of Washington Department of Conservation and Development, 1943.

List of Geological Maps Published by the Quebec Department of Mines. Quebec, 1944.

Lower Romaine River Area, Saguenay County. By J. A. Retty. Geological Report 19, Division of Geological Surveys. Quebec, 1944.

McLouth Gas and Oil Fields, Jefferson and Leavenworth Counties, Kansas. By Wallace Lee and Thomas G. Payne. Bulletin 53, State Geological Survey of Kansas. Lawrence, 1944.

The Mammals of the Vernay-Hopwood Chindroin Expedition, Northern Burma. By T. Donald Carter. Bulletin of the American Museum of Natural History, Vol. LXXXII, Art. 4. New York, 1944.

Man and His Physical World. By Dwight E. Gray. New York: D. Van Nostrand Co., 1942.

Manganiferous Iron-Ore Deposits near Silver City, New Mexico. By Lawson P. Entwistle. New Mexico School of Mines Bulletin No. 19. Socorro, N.M., 1944.

Man's Physical Universe: A Survey of Physical

Science for Colleges. By Arthur Talbot Bawden: New York: Macmillan Co., 1943.

Mined Areas of the Weir-Pittsburg Coal Bed. By G. E. Abernathy. Bulletin 52, State Geological Survey of Kansas. Lawrence, 1944.

Mineral Resources of South Carolina. By the South Carolina State Planning Board. Columbia, 1944.

The Mining Industry of the Province of Quebec. Department of Mines. Quebec: Redempti Paradis, 1943.

Mississippi Minerals. By William Clifford Morse. Mississippi State Geological Survey Bulletin 59. University, Miss., 1944.

Monroe County Mineral Resources. Geology by Franklin Earl Vestal; tests by Thomas Edwin McCutcheon. Mississippi State Geological Survey Bulletin 57. University, Miss., 1943.

A National Oil Policy for the United States Petroleum Industry War Council. Washington, D.C.: Superintendent of Documents, 1944.

A New Fossil Whale from the Miocene of Peru. By Edwin H. Colbert. Bulletin of the American Museum of Natural History, Vol. LXXXIII, Art. 3. New York, 1944.

Nineteenth Biennial Report, July 1, 1942, to June 30, 1944, Mississippi Geological Survey. By William Clifford Morse. University, Miss., 1943.

Oil and Gas Development in Illinois in 1943. By Alfred H. Bell and Charles W. Carter. Press Bulletin Series No. 50, Illinois State Geological Survey. Urbana, 1944.

The Opaoka River Area: Abitibi Territory. By B. C. Freeman and J. M. Black. Geological Report 16. Quebec: Redempti Paradis, 1944.

The Organization of the New York Academy of Sciences, Revised to August 1, 1944. By Eunice Thomas Miner. Annals of the New York Academy of Sciences, Vol. XLV, Art. 8. New York, 1944.

Palaeontology of Harrar Province, Thipoia. Part I: The Dudley Expedition. By Barnum Brown. Part II: Echinoidea. By Ethel D. Currie. Part III: Jurassic Anthozoa and Hydrozoa. By John W. Wells. Part IV: Jurassic Cephalopoda and a Cretaceous Nautilus. By Gayle Scott. Bulletin of the American Museum of Natural History, Vol. LXXXII, Arts. 1, 2, and 3. New York, 1943.

Physical Science. By William F. Ehret (ed.). New York: Macmillan Co., 1942.

Preliminary Report of High Alumina Iron Ores in Washington County, Oregon. By F. W. Libbey, W. D. Lowry, and R. S. Mason. G.M.I. Short Paper No. 12. Portland, Ore.: Department of Geology and Mineral Industries, 1944.

A Process for Extracting Alumina from Kansas Clay. By E. D. Kinney. Bulletin 47, Part IV, State Geological Survey of Kansas. Lawrence, 1943.

Progress Reports on Subsurface Studies of the Pennsylvania System in the Illinois Basin. Report of Investigations No. 93, Illinois State Geological Survey. Urbana, 1944.

Rare-Element Minerals in Pegmatites, Yellowknife-Beaulieu Area, Northwest Territories. By A. W. Jolliffe. Geological Survey Paper 44-12. Ottawa, Canada; Department of Mines and Resources, Mines and Geology Branch, 1944.

Reconnaissance of Pleistocene Deposits in North-Central Kansas. By Claude W. Hibbard, John C. Frye, and A. Byron Leonard. Bulletin 52, Part I, State Geological Survey of Kansas. Lawrence, 1944.

Report of the Committee on Marine Ecology as Related to Paleontology, 1942–1943. Appendix Q: Annual Report of the Division for 1942–1943. Washington, D.C.: National Research Council, 1943.

Report of the Committee on Marine Ecology as Related to Paleontology, 1943–1944. Washington, D.C.: National Research Council, 1944.

Report of the Committee on the Measurement of Geologic Time, 1942–1943. Appendix F: Annual Report of the Division with Exhibits A–E. Washington, D.C.: National Research Council, 1943.

Report of the Geological Survey for the Year 1940. Perth, Western Australia, 1941.

Report Submitted to the Trail Smelter Arbitral Tribunal. By R. S. Dean and R. E. Swain. U.S. Bureau of Mines Bulletin 453. Washington, D.C.: Government Printing Office, 1944.

Sillimanite Deposits in the Monadnock Quadrangle, New Hampshire. By Katharine Fowler-Billings. Concord: New Hampshire State Planning and Development Commission, 1944.

Soil and Plant Analysis. By C. S. Piper. New York: Interscience Publishers, 1944.

Some Granitic Rocks of South-eastern South Australia. By D. Mawson and L. W. Parkin.

Transactions of the Royal Society of South Australia, Vol. LXVII (1943).

Stratigraphic Aspects of the Blewett–Cle Elum Iron Ore Zone, Chelan and Kittitas Counties, Washington. By R. L. Lupher. Report of Investigations No. 11. Olympia: State of Washington Department of Conservation and Development, 1944.

Stratigraphy of the Colorado Group, Upper Cretaceous, in Northern New Mexico. By Charles H. Rankin. State Bureau of Mines and Mineral Resources Bulletin No. 20. Socorro: New Mexico School of Mines, 1944.

The Stratigraphy of Some Lower Ordovician Formations of the Ozark Uplift. By James S. Cullison. University of Missouri, School of Mines and Metallurgy, Technical Series Bulletin, Vol. XV, No. 2. Rolla, 1944.

The Stratigraphy and Structural Development of the Forest City Basin in Kansas. By Wallace Lee. Bulletin 51, State Geological Survey of Kansas. Lawrence, 1943.

The Structural Character of the Flinders Ranges. By D. Mawson. Transactions of the Royal Society of South Australia, Vol. LXVI (1942).

A Study of the Troödont Dinosaurs with the Description of a New Genus and Four New Species. By B. Brown and Erich M. Schlaikjer. Bulletin of the American Museum of Natural History, Vol. LXXXII, Art. 5. New York, 1943.

Summarized Records of Deep Wells. By R. C. Tucker. West Virginia Geological Survey, Vol. XVI. Charleston, 1943.

Tabular Description of Outcropping Rocks in Kansas. By Raymond C. Moore, John C. Frye, and John Mark Jewett. Bulletin 52, State Geological Survey of Kansas. Lawrence, 1944.

Transactions of the American Geophysical Union, 1944. Part I: Reports and Papers of the Joint Regional Meeting Section of Hydrology, Western-Snow Conference, American Society of Agricultural Engineers, Berkeley, California. Part II: Reports and Papers, Section of Hydrology. Part III: Reports and Papers of the Joint Regional Meeting, Section of Hydrology, Western-Snow Conference, Corvallis, Oregon.

United States Foreign Oil Policy and Petroleum Reserves Corporation. Washington, D.C.: Petroleum Industry War Council, 1944.

Volcanoes of the Three Sisters Region, Oregon Cascades. By Howel Williams. Berkeley and Los Angeles: University of California Press, 1944.

Wyoming Vermiculite Deposits. By Arthur F. Hagner. Geological Survey of Wyoming Bulletin No. 34. Laramie: University of Wyoming, 1944.

MAP RECEIVED

Mapa geologico generalizado de Bolivia. By Victor Oppenheim. Scale 1/2,000,000. Lima: Geological Society of Peru, 1944.

VOLUME LIII NUMBER 2

THE JOURNAL OF GEOLOGY

March 1945

PROBLEMS OF PLEISTOCENE STRATIGRAPHY IN CENTRAL AND WESTERN KANSAS

JOHN C. FRYE
State Geological Survey, University of Kansas, Lawrence

ABSTRACT

The nonglacial Pleistocene formations of central and western Kansas have been studied at many localities during the last ten years. In this paper the results of these investigations are reviewed, the major aspects of the stratigraphy are discussed, and the most significant problems are listed. Recent work has resulted in the assignment of a Pleistocene age to heterogeneous deposits, formerly believed to be Tertiary, that occur widely in western Kansas and the assignment of a Tertiary age to some sediments of central Kansas, formerly supposed to be Pleistocene, that now are differentiated as the Emma Creek formation. Some perplexing problems have been answered, but new and formerly unsuspected problems have been brought to light. The most important stratigraphic problem is the proper definition and delineation of the Pliocene-Pleistocene boundary. Other problems that warrant detailed future studies are correlations among the several stratigraphic units discussed and correlation of these nonglacial formations with the glacial formations of the upper Mississippi Valley.

INTRODUCTION

Studies of the nonglacial Pleistocene deposits of central and western Kansas have been revived after nearly half a century of inactivity, and during the past several years they have been carried on extensively. M. K. Elias[1] in 1931 brought to notice some Pleistocene problems in writing on the geology of Wallace County. Real impetus to work on the western Kansas Pleistocene deposits has been given by C. W. Hibbard, who has collected and described fossil mammals from many localities. In 1938 the State Geological Survey of Kansas and the United States Geological Survey initiated regional co-operative ground-water studies over many of the southwestern Kansas counties. Because the surficial water-bearing rocks of western Kansas are largely of Pliocene and Pleistocene age, this program led to detailed county studies of these deposits. As a preliminary step in this program, the regional geology of southwestern Kansas was studied by H. T. U. Smith.[2] Ground-water studies have been made in central and western Kansas by Frank Byrne, V. C. Fishel, John C. Frye, Bruce F. Latta, Stanley W. Lohman, Thad G. McLaughlin, H. A. Waite, and C. C. Williams, all members of the state or federal geological survey staffs. Recently, A. Byron Leonard has made detailed studies of the molluscan faunas. The published works of these men have been used extensively in the preparation of this paper.

[1] "The Geology of Wallace County, Kansas," *Kan. Geol. Surv. Bull. 18* (1931), pp. 1–254.

[2] "Geologic Studies in Southwestern Kansas," *Kan. Geol. Surv. Bull. 34* (1940), pp. 1–240.

Central and western Kansas is situated in the High Plains and Plains Border sections of the Great Plains physiographic province and the western part of the Osage Plains section of the Central Lowlands physiographic province.[3] Most of the central Great Plains region was subjected to subaerial erosion during early Tertiary time. Cenozoic deposition was initiated at different times in the various parts of the region. In northern Nebraska, earliest Tertiary deposition occurred during the Oligocene epoch[4] and is represented by the Chadron formation and the Brule clay. These were followed, in Miocene time, by the deposition of beds comprising the Arikaree and Hemingford groups, and in Pliocene time by the Ogallala group. South of Kansas, in western Oklahoma, the oldest known Tertiary formation is the Laverne[5] of early Pliocene and possibly latest Miocene age. Tertiary deposits older than Pliocene are not known to exist in Kansas; but, with possible exception of an area in the west-central part of the state, the High Plains region of Kansas was completely covered by stream deposits at some time during the Pliocene. Extensive Pleistocene deposits occur in central and western Kansas; they lie unconformably on Tertiary deposits, as basin fillings, terrace deposits, abandoned valley fillings, and upland deposits of loess and dune sand.

The purpose of this paper is to summarize the major aspects of the stratigraphy of the Pleistocene deposits in the Great Plains of Kansas and to point out some of the most significant problems so far recognized.

GENERAL CONSIDERATIONS

DEFINITION OF PLEISTOCENE

In any general consideration of the Pleistocene problems of central and western Kansas, the criteria used in determining the boundary line between beds of Pliocene and Pleistocene age assume great importance. In order better to understand the problems in the Great Plains, it seems advisable to review briefly the development of the term "Pleistocene."

General.—Subdivisions of the Cenozoic era commonly used by geologists were originally based on the percentage of living species found to occur as fossils among the marine molluscan faunas of the various strata. Charles Lyell[6] in 1839 first used the term "Pleistocene" to designate those marine strata in the Mediterranean region which contain 70 per cent or more of modern species. In 1846 Edward Forbes[7] considered Lyell's term "Pleistocene" a synonym for "glacial beds" and applied it to the glacial deposits in the British Isles. In 1873 Lyell[8] referred to Forbes's usage of "Pleistocene" and stated that it had not been used as short for "Newer Pliocene," as Lyell had originally intended, but as "Post-Pliocene." Lyell advised that the

[3] N. M. Fenneman, "Physical Divisions of the United States (Map)," *U.S. Geol. Surv.* (1930), scale 1:7,000,000.

[4] A. L. Lugn, "Classification of the Tertiary System in Nebraska," *Bull. Geol. Soc. Amer.*, Vol. L (1939), p. 1264.

[5] John C. Frye and Claude W. Hibbard, "Pliocene and Pleistocene Stratigraphy and Paleontology of the Meade Basin, Southwestern Kansas," *Kan. Geol. Surv. Bull. 38*, Part XIII (1941), pp. 398–403.

[6] *Elements of Geology* (French translation, Paris, 1839), Appendix, pp. 616–21.

[7] "On the Connexion between the Distribution of the Existing Fauna and Flora of the British Isles, and the Geological Changes Which Have Affected Their Area, Especially during the Epoch of the Northern Drift," *Great Britain Geol. Surv. Mem.* 1 (1846), pp. 402, 403.

[8] *Antiquity of Man* (4th ed.; London: Murry, 1873), pp. 3, 4.

term "Pleistocene" be used in a stratigraphic sense to apply to the lower subdivision of the Post-Tertiary (Post-Pliocene).

J. E. Eaton[9] has critically discussed the early usage of "Pleistocene" and has pointed out that Lyell's 1873 restriction reduced the average extinction of marine molluscan species to 10 per cent or less. He also suggested that Lyell's 1873 usage of "Pleistocene" in a purely stratigraphic sense, although including all the glacial section of the British Isles, did not necessarily imply age correlation of any of the glacial beds with any of the typical marine beds of the Mediterranean region.

M. Grace Wilmarth,[10] in 1925, reported acceptance by the United States Geological Survey of Lyell's revised definition of Pleistocene and stated that it "includes the deposits of the Great Ice Age, as it is popularly known, and contemporaneous marine, fluviatile, lacustrine, and volcanic rocks. In some areas it also probably includes some preglacial deposits and some post-glacial deposits older than those of the Recent epoch."

Eaton[11] stated that deposits classed as Lower (marine) Pleistocene in California are certainly older than the oldest known glacial deposits of the state and that the seas—although cooler than during the preceding Pliocene time—were not cold or glacial, as would be expected if the beginning of Pleistocene time were synchronous with a widespread continental glaciation. It is possible, as he implies for California, that minor ice advances may have occurred prior to the oldest recorded glaciation, their deposits being completely destroyed by a more widespread glaciation.

Glacial Pleistocene deposits of North America have been studied in detail in areas where they do not interfinger with marine beds or overlie marine Upper Pliocene deposits. Possibly for this reason, a more or less distinct definition of the Pleistocene has been used in the glaciated areas and has been generally accepted by many geologists as the standard definition of the series. In central North America, Pleistocene time is considered to have started with the advance of the Nebraskan continental ice sheet. This might establish a fairly satisfactory time-marker, but there does not seem to be consistency in the thinking of various geologists as to whether the Pleistocene began (*a*) when the ice fields started to form at the various centers of accumulation, (*b*) after the ice had moved out a little way from these centers, (*c*) when the ice crossed the international boundary line, or (*d*) when it reached its maximum extent. Although it is not uncommon for a lithologically distinct formational unit, such as a till sheet, to be of slightly different ages at different places, a unit of time, such as an epoch, cannot well have a sliding scale at its base. If the point in time which marks the first accumulation of ice that gave rise to the first major ice sheet (the Nebraskan) is arbitrarily chosen, we have an understandable definition for the beginning of Pleistocene time in interior continental North America; however, we do not know whether such an arbitrary time line coincides with the time line or lines used in marine sections

[9] "The Pleistocene of California," in "Geologic Formations and Economic Development of the Oil and Gas Fields of California," prepared under the direction of Olaf P. Jenkins, *Calif. Dept. of Natural Resources, Div. of Mines Bull. 118* (1943), pp. 203–4.

[10] "The Geologic Time Classification of the United States Geological Survey Compared with Other Classifications," *U.S. Geol. Surv. Bull. 769* (1925), p. 49.

[11] Ftn. 9 (1943).

or in glacial sections of other parts of the world, and we do not know whether satisfactory correlation of these various lines can be established.

Central Great Plains.—Still another definition of Pleistocene time has been used as applied to the nonmarine-nonglacial deposits of the central Great Plains. This generally used but mostly unstated definition may or may not coincide with either of the other two, but it is probably more closely related to glacial than to marine usage.

In a committee report published in 1941,[12] mammals considered to be characteristic of Pleistocene time were listed, and the entire assemblage was referred to the *Equus* faunal zone. It is seemingly impossible, however, to specify a vertebrate fauna that can be used as an index to the lowermost Pleistocene over all of the western interior of the United States. Many species of vertebrates range vertically from Pliocene to Pleistocene. Mammals, the most abundant and most diagnostic of the vertebrates, migrate readily in response to climatic change. The geographic range of many mammalian species probably shifted entirely across the United States more than once during Pleistocene time, and now similar assemblages of fossil mammals can be found in the High Plains deposits separated vertically by a different assemblage of fossil mammals. The migratory nature of these forms leads one to expect that the Aftonian fauna of southern Canada should possess more forms in common with the Nebraskan fauna than with the Aftonian fauna of southern United States. The foregoing considerations suggest that it is not possible to make certain and detailed correlations of Pleistocene beds within the Great Plains region by the simple expedient of matching fragmentary vertebrate faunas. In order that detailed correlations have much validity, they must be based on relatively large faunal assemblages and must be consistent with the climatic history of the region.

In western Kansas, in the Meade Basin only, beds have been studied that are known to be older than the basal Meade gravels, classed as Pleistocene, and younger than the "cap rock" or "Algal limestone" that marks the top of the Middle Pliocene Ogallala. It is in this area, therefore, that the problem of defining the Lower Pleistocene boundary assumes the most significant importance. For many years most of the beds in this structural basin, including the Meade formation and Kingsdown silt, were referred to the Tertiary, although S. W. Williston[13] in 1897 pointed out that the fossils collected by Cragin should be considered Pleistocene. During the past decade, Hibbard has described vertebrate fossils, from the strata above the Rexroad member of the Ogallala, that he believed were Pleistocene; and until the publication in 1944 of a paper by Paul O. McGrew[14] the usage generally accepted was that of Frye and Hibbard[15] in their discussion of the stratigraphy and paleontology of that area. The determination of the beginning of Pleistocene time was based primarily on fossil vertebrates and lithologic evidence that seems to indicate a widespread break in sedimentation, followed by the deposi-

[12] Horace E. Wood *et al.*, "Nomenclature and Correlation of the North American Continental Tertiary," *Bull. Geol. Soc. Amer.*, Vol. LII (1941), p. 13, Pl. 1.

[13] "The Pleistocene of Kansas," *Kan. Univ. Geol. Surv.*, Vol. II (1897), p. 300.

[14] "An Early Pleistocene (Blancan) Fauna from Nebraska," *Field Mus. Nat. Hist., Geol. Ser.*, Vol. IX, No. 2 (1944), pp. 33–66.

[15] Pp. 389–424 of ftn. 5 (1941).

tion of sediments having coarser texture than the underlying Tertiary sediments. These coarse sediments may be the result of the first glaciation in the Rocky Mountain region to the west. The contact, however, is not everywhere overlain by coarse gravels, and the strata classed as Pleistocene contain a large percentage of fine-textured deposits. In many places this disconformity merely serves to localize the contact between beds that bear two vertebrate faunas, the fossil vertebrates actually constituting the dating criteria.

McGrew[16] correlated the fauna of the supposed Upper Pliocene Rexroad member of the Ogallala formation with his Sand Draw fauna of Nebraska which he believes to be of Pleistocene (Aftonian) age. This correlation again raises the problem of delineation of the Pliocene-Pleistocene boundary line. Although the exact age of the Rexroad beds is now in doubt, there are several aspects of McGrew's basis of correlation that lead the present writer to believe that the earlier dating, by Frye and Hibbard, of the Rexroad member as Upper Pliocene is more consistent with the definition of Pleistocene time as used in the upper Mississippi Valley. It should be pointed out that three of the mammals listed by McGrew[17] as occurring in the Rexroad fauna were earlier questioned as being truly a part of that fauna,[18] and subsequent field observations have strengthened the belief that these forms actually came from overlying Pleistocene deposits. The vertebrate fauna of the Rexroad is a warm-climate fauna and was taken in its entirety from the uppermost 35–40 feet of a member known to have a total thickness of about 250 feet. A. Byron Leonard, of the University of Kansas, is now studying snails from test holes drilled through the entire thickness of the Rexroad. These snails were taken from samples at depths of as much as 200 feet below the top of the member, and they also indicate a warm climate. Stratigraphically, the Rexroad beds are believed to rest conformably upon beds generally considered to be of Middle Pliocene age, and they are separated from the overlying Meade formation by a distinct unconformity. This unconformity can be seen at the type locality of the Rexroad and along canyons for 10 miles north of that locality; but at the type locality of the Meade formation, although the Rexroad beds are believed to be absent, the unconformity at the base of the Meade formation is less clearly recognizable in the field. Hibbard[19] has correlated a cold-climate fauna (Cudahy), that occurs stratigraphically more than 30 feet above the base of the Meade formation and below a warm-climate fauna (Borchers), with the fauna of the oldest terrace deposit along the Smoky Hill Valley. This old terrace is the highest and oldest of three distinct Pleistocene terraces and is probably continuous with the other extensive high terraces of northern Kansas. As Kansan and Nebraskan tills are both known to occur in northeastern Kansas, it seems likely that terraces would have developed during Nebraskan time if extensive terraces were formed in the area in Kansan time. Owing to the incomplete nature of the field data, however, definite correlations by this method must await more extensive and detailed studies. Regardless of the validity of McGrew's

[16] Ftn. 14 (1944).

[17] *Ibid.*, p. 39.

[18] Frye and Hibbard, p. 409 of ftn. 5 (1941).

[19] "Stratigraphy and Vertebrate Paleontology of Pleistocene Deposits of Southwestern Kansas," *Bull. Geol. Soc. Amer.*, Vol. LV (1944), pp. 707–54.

faunal correlations, his conclusion that certain fossil mammals of the plains region represent the Aftonian of the glacial section[20] is in part based on unpublished geological observations indicating that deposits at the Broadwater locality in Nebraska are of Aftonian age.

McGrew[21] states that there is little geological evidence for assigning the Rexroad to the late Pliocene. However, the geological evidence is not contradictory to such a classification because these beds seem to be generally conformable on known Pliocene and unconformable below deposits that are agreed to belong to the Pleistocene. They contain only a small percentage of material that has the appearance of redeposited Ogallala, whereas the overlying beds contain abundant abraded pebbles of "mortar beds" and caliche. The Rexroad is restricted to a structural "trap" produced by faulting, whereas the Pleistocene deposits of the area occur widely beyond the limits of such "traps." It is hardly to be expected that sediments accumulated in central and western Kansas only in one small structural "trap" during Nebraskan and Aftonian time and were deposited throughout much of the region during Kansan and Yarmouth time. Although movement along faults was initiated in Rexroad time, it continued into definite Pleistocene time, and most of the major sinkhole development occurred subsequent to deposition of the Rexroad.

The definition of the beginning of Pleistocene time in central and western Kansas should be considered as yet an unsolved problem. For the purpose of this paper Frye and Hibbard's usage is followed, and in the Meade Basin area only those beds lying stratigraphically above the Rexroad member of the Ogallala formation, and generally agreed to be Pleistocene, are included within the Pleistocene.

Owing to the scarcity of Pleistocene fossil vertebrates in the upper Mississippi Valley region, it will probably be impossible to make satisfactory correlations between the central Great Plains and the glaciated areas on the basis of fossils alone. It seems likely that the most feasible method of detailed correlations between these two regions is to trace terrace deposits eastward down the major drainage ways into the glaciated area.

STRATIGRAPHIC USAGE

The classification of the nonglacial Pleistocene beds in the central Great Plains has not been governed clearly in all cases by the rules of stratigraphic nomenclature.[22] There is uncertainty inasmuch as many Pleistocene deposits in this area that underlie terraces at several levels (now partly dissected and discontinuous), fill isolated sinkholes, and occur on the bottoms of abandoned high level valleys, may be equivalent in age to various parts of a more widespread formation deposited in an extensive basin or on the uplands. Article 18 of the "Rules of Stratigraphic Nomenclature"[23] states that "formal names shall not be applied to deposits of merely local extent." The Wisconsin drift is cited as an example of a widespread Pleistocene unit that should be governed by the rules applying to formal names. The only Pleistocene unit in central and western Kansas comparable in lithologic uniformity and geographic extent to the Wisconsin drift is the upper gray loess of the Sanborn formation. The problem at

[20] McGrew, p. 37 of ftn. 14 (1944).

[21] *Ibid.*, p. 38.

[22] G. H. Ashley *et al.*, "Classification and Nomenclature of Rock Units," *Bull. Geol. Soc. Amer.*, Vol. XLIV (1933), pp. 423–59.

[23] *Ibid.*, p. 441.

once arises as to whether local deposits having a certain uniformity of lithologic character should be referred to an adjacent more widespread formation, representing somewhat dissimilar environmental conditions and perhaps a greater span of time, or, on the other hand, severally should be given formal stratigraphic names.

Local names are now applied to a number of slightly dissimilar and disconnected deposits in central and western Kansas, and there are unnamed minor and isolated deposits. It may be found advisable at some future date to apply certain of the existing names more widely and to discontinue the local names now in use. At present, however, it seems that the retention of local names, each of which has a specific meaning and usage somewhat different from any other, results in greater clarity and usefulness in the Pleistocene classification.

PREVIOUS WORK ON KANSAS PLEISTOCENE DEPOSITS

The presence of glacial deposits in northeastern Kansas has been known for more than fifty years,[24] and, although there has been some question as to the correlation of Nebraskan and Aftonian deposits in this area,[25] glacial till has been recognized with certainty over most of the area north of the Kansas River Valley and east of central Washington County.[26]

Pleistocene deposits in central Kansas were first described in 1892 and referred to the "*Equus* beds" by J. Lindahl.[27] The age of these deposits, which were also early referred to as the "Sheridan beds,"[28] was based on the fossil vertebrates collected from several sand and gravel pits, particularly from a pit in northwestern McPherson County now believed to occur on the west side of the filled and abandoned McPherson Valley. These beds subsequently were studied in some detail by Erasmus Haworth and J. W. Beede,[29] who in 1897 renamed them the "McPherson *Equus* beds."

The Pleistocene of northwestern Kansas has been almost completely unknown until recent years, although Robert Hay's[30] "Tertiary marl" was believed to extend over that part of the state. An early reference to Pleistocene terrace deposits in Russell County was made in 1897 by W. N. Logan,[31] who named the "Salt Creek gravel beds." Elias[32] first

[24] G. C. Swallow, "Quaternary Deposits of Missouri," *Proc. Amer. Assoc. Adv. Sci.*, Vol. XI, Part II (1858), pp. 21–39; B. F. Mudge, "First Annual Report of Geology of Kansas" (1866), pp. 12–15; T. C. Chamberlin, "Preliminary Paper on the Terminal Moraine of the Second Glacial Epoch," *Ann. Rept. U.S. Geol. Surv.*, Vol. III (1883), map, pp. 291–402; Chamberlin, in James Geikie's *The Great Ice Age* (New York: D. Appleton & Co., 1895), pp. 724–75; J. E. Todd, "Kansas during the Ice Age," *Kan. Acad. Sci. Trans.*, Vol. XXVIII (1918), pp. 33–47; W. H. Schoewe, "Glacial Geology of Kansas," *Pan-Amer. Geol.*, Vol. XL (1923), pp. 102–10; G. F. Kay and Earl T. Apfel, "The Pre-Illinoian Pleistocene Geology of Iowa," *Iowa Geol. Surv.*, Vol. XXXIV (1928), pp. 1–304; W. H. Schoewe, "Evidence for a Relocation of the Drift Border in Eastern Kansas," *Jour. Geol.*, Vol. XXXVIII (1930), pp. 67–77.

[25] Frye, "Reconnaissance of Ground-Water Resources in Atchison County," *Kan. Geol. Surv., Bull. 38*, Part IX (1941), pp. 246–47.

[26] Schoewe, "Evidence for the Relocation of West Drift Border in Eastern Kansas," *Kan. Acad. Sci. Trans.*, Vol. XLII (1939), p. 367.

[27] "Description of a Skull of *Megalonyx leidyi*, n. sp.," *Trans. Amer. Phil. Soc.*, Vol. XVII (1892), pp. 1–10.

[28] W. B. Scott, *Introduction to Geology* (New York: Macmillan Co., 1897), pp. 532–33.

[29] "The McPherson *Equus* Beds," *Kan. Univ. Geol. Surv.*, Vol. II (1897), pp. 285–96.

[30] *Kan. State Board of Agric., 8th Biennial Report* (1893), p. 101.

[31] "The Upper Cretaceous," *Kan. Univ. Geol. Surv.*, Vol. II (1897), pp. 218–19.

[32] Ftn. 1 (1931).

described in 1931 the occurrence in northwestern Kansas of Pleistocene loess and scattered gravel deposits, which he named the "Sanborn formation."

For many years Pleistocene fossils have been known to occur in southwestern Kansas, but no distinct formational units of Pleistocene age were defined prior to the present decade. In 1893 Hay[33] named and described deposits now known to be of Pleistocene age as the "Plains marl" or "Tertiary marl." Later, F. W. Cragin[34] collected fossils from Pleistocene deposits and in 1896 named the "Meade gravels," "Pearlette ash," and "Kingsdown marl"; however, he believed that these deposits were Pliocene rather than Pleistocene in age. Williston[35] in 1897 stated that the fossils listed by Cragin as occurring in three supposed Pliocene formations clearly showed them to be of Pleistocene age. However, he questioned the great thickness of Pleistocene deposits thus implied. At the same time, Williston[36] stated that the "Plains marl" was evidently a part of the "*Equus* beds." In 1896 and 1897, Haworth[37] published data concerning the structure and ground-water resources of beds in southwestern Kansas now classed as Pleistocene. In 1901 W. D. Johnson[38] described in great detail the unconsolidated deposits of the southern High Plains; but, as he did not subdivide these beds or specifically designate a Pleistocene formation, he presumably believed them to be largely of Tertiary age.

AREAS OF PLEISTOCENE DEPOSITS IN KANSAS

The Pleistocene deposits of Kansas are discussed in this paper by geographic regions: (1) the northeastern Kansas area that was subjected to continental glaciation, which is now covered by till, glacio-fluvial, and eolian deposits; (2) the large area in central, north-central, and northwestern Kansas that contains extensive nonglacial deposits consisting mostly of valley fills (including high-level abandoned valleys), terrace deposits, and extensive and widespread deposits of eolian loess that mantle both the water-laid Pleistocene deposits and the older rocks; (3) the southwestern Kansas area that is characterized by thick, more or less persistent, basin deposits and associated terrace beds; and (4) the southeastern Kansas area that contains thin, nonpersistent terrace deposits, generally unrelated to the problems discussed in this paper.

CENTRAL KANSAS

During the last sixty years the "*Equus* beds area" in Sedgwick, Harvey, McPherson, and adjacent counties of central Kansas has been a profitable collecting ground for Pleistocene fossil mammals. Haworth and Beede[39] considered all the unconsolidated deposits lying above the Permian in this area as part of the "McPherson *Equus* beds" of Pleistocene age. They believed all the deposits to be of fluvial origin, but, owing to the height of the "McPherson ridge" and the

[33] Ftn. 30 (1893).

[34] "Preliminary Notice of Three Late Neocene Terranes," *Colorado College Studies*, Vol. VI (1896), pp. 53-54.

[35] Ftn. 13 (1897).

[36] *Ibid.*, p. 306.

[37] "Local Deformation of Strata in Meade County, Kansas, and Adjoining Territory," *Amer. Jour. Sci.*, Vol. II (4th ser., 1896), pp. 368-73; "Underground Waters of Southwestern Kansas," *U.S. Geol. Surv., Water Supply Paper 6* (1897), pp. 1-65; "Physiography of Western Kansas," *Kan. Univ. Geol. Surv.*, Vol. II (1897), pp. 11-49.

[38] "The High Plains and Their Utilization," *U.S. Geol. Surv., 21st Ann. Rept.*, Part IV (1901), pp. 601-741.

[39] Ftn. 29 (1897).

supposed absence of terraces, they stated that they were at a loss to explain the history of the depositing stream. In 1937 a study of ground-water resources of this area was undertaken by the state and federal geological surveys, working in co-operation. As a result of these studies an extensive body of eolian loess, believed to attain a maximum thickness of more than 100 feet under the "McPherson ridge,"[40] was separated from the underlying water-laid Pleistocene deposits. Evidence was discovered also indicating that the unconsolidated stream deposits capping the uplands are of Pliocene rather than Pleistocene age.[41] Accordingly, the term "McPherson formation" was restricted to include only the fluvial Pleistocene deposits, and a new formation, the Emma Creek, was named and described to include the newly differentiated Pliocene deposits. The stratigraphy and vertebrate fauna of these two formations were discussed in more detail in 1941.[42] Deposits that are believed to have been continuous with the restricted McPherson formation recently have been traced westward as terrace deposits along the Smoky Hill River Valley.[43] Isolated areas of deposits of comparable Pleistocene age are known to occur on intermediate topographic levels north of the "*Equus* beds area" in Saline, Ottawa, and Cloud counties.

[40] Frye, "Physiographic Significance of Loess Near McPherson, Kansas," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXIII, No. 8 (1939), pp. 1232–33.

[41] Stanley W. Lohman and John C. Frye, "Geology and Ground-Water Resources of the 'Equus Beds' Area in South-Central Kansas," *Econ. Geol.*, Vol. XXXV (1940), pp. 839–66.

[42] Frye and Hibbard, "Stratigraphy and Paleontology of a New Middle and Upper Pliocene Formation of South-Central Kansas," *Jour. Geol.*, Vol. XLIX, No. 3 (1941), pp. 261–78.

[43] John C. Frye, A. Byron Leonard, and Claude W. Hibbard, "Westward Extension of the Kansas 'Equus Beds,'" *Jour. Geol.*, Vol. LI, No. 1 (1943), pp. 33–47.

CHARACTER OF THE DEPOSITS

The Pleistocene deposits of central Kansas consist of gravel, sand, and silt, which occur as abandoned valley fillings and terrace deposits, and extensive areas of eolian loess that blankets both the older Pleistocene beds and the pre-Quaternary rocks. Volcanic ash deposits have been observed that are believed to be of two distinct ages. Pleistocene stream deposits also occur as the older part of the alluvial fill along such valleys as the Arkansas, although the alluvium below the flood plains of other streams, such as the Smoky Hill, is believed to be entirely Recent in age.

The dating of the several unconsolidated formations in central Kansas has been based largely on fossil mammals, although the various stratigraphic units have been differentiated in the field by physical means. Haworth and Beede described these unconsolidated deposits in the "*Equus* beds" area as being free from arkosic materials; Lohman and Frye stated that the Pleistocene deposits contain appreciable percentages of feldspar and granite grains, although the Emma Creek is virtually free from them. Subsequent field observations have revealed that locally the sand and gravel beds of the Emma Creek formation (particularly the "lower" Emma Creek, or Middle Pliocene part) are equally as arkosic as the Pleistocene deposits. In northern McPherson County and northward the two formations can generally be separated by their topographic position. The Pleistocene stream deposits occur as valley fillings and on terrace levels, whereas the Tertiary deposits universally occur as a capping upland formation. Either formation may be overlain by loess.

MAJOR PROBLEMS

Several aspects of the stratigraphy of the Pleistocene beds in central Kansas, are not adequately known. Throughout the area of Tertiary deposits, scattered Pleistocene fossils have been collected along the minor stream valleys. At a few places thin terrace deposits overlying Tertiary beds are exposed. Farther east, particularly along the headwater streams of Walnut River, thin Pleistocene terrace deposits have been found resting directly on Permian bedrock. The exact correlation and regional significance of these minor Pleistocene deposits are yet to be determined.

Several terraces of different ages within the Pleistocene have been described along Smoky Hill Valley northwest of the "*Equus* beds area." The stream that deposited the oldest of these terrace beds most certainly joined the stream that flowed southward through the now abandoned McPherson Valley; therefore, the deposits on one or more of these terrace levels are equivalent to the McPherson formation that fills the large north-south abandoned valley. It is yet to be determined whether the major valley filling represents the time equivalent to the oldest or the intermediate of these terraces farther west, or whether it includes beds of both ages. Because this filled valley lies close to the eastern edge of the Hutchinson salt bed of the Permian Wellington formation (less than 300 feet below the bedrock floor), it probably has been modified by solution and subsidence; therefore, physiographic position, a criterion that generally serves to differentiate the Pliocene from the Pleistocene in north-central Kansas, cannot be used safely to distinguish minor age differences within the Pleistocene.

Knowledge of the vertebrate faunas is not complete enough as yet to allow detailed correlations. These considerations serve to obscure the exact age of the restricted McPherson formation within the Lower Pleistocene, and which, or how many, of the important terrace beds were deposited contemporaneously with the beds at the type locality of the McPherson formation. It is the opinion of the writer, based on relations of terraces to the till sheets to the northeast and of the McPherson to the younger loess, that all the beds referred to the McPherson formation are pre-Illinoian, but data are not yet available to make this opinion more than a working hypothesis.

A prominent disconformity occurs within the Pliocene Emma Creek formation. The beds below this disconformity have yielded a characteristic Middle Pliocene vertebrate fauna,[44] but the beds above it have yielded only few fragmentary fossils; and, although physiographic data conclusively show the upper part of the Emma Creek to be older than the McPherson formation, it does not exclude the possibility that the formation may include some beds of earliest Pleistocene age.

The upper surface of the Emma Creek formation, where it is well developed in east-central McPherson County, underlies the surface of a relatively smooth and accordant upland plain. This upland plain slopes gently to the south-southwest and can be traced, seemingly without a break, southward to Wichita. This surface extends south from Wichita along the east side of the Arkansas Valley, and southwest from Wichita it is seemingly accordant with the upland plain that extends south and southwest from the Arkansas Valley to the Oklahoma line. This topography leads one to believe that the deposits underlying such

[44] Personal communication from Claude W. Hibbard.

a seemingly accordant surface are all of similar age, namely, Pliocene. Unfortunately for such a hypothesis, a relatively large number of fossil mammals of Middle to late Pleistocene age have been collected from the deposits underlying this surface south of Arkansas River.[45] Remnants of a surface that has the appearance of being a continuation of this central Kansas plain extend westward for nearly 200 miles north of the Oklahoma line. This is not a smooth upland surface such as occurs in McPherson County, but is more discontinuous, broken by areas of Permian bedrock exposures, and has the appearance of a hummocky or undulating plain. The post-Permian deposits under this surface for the entire distance are Upper Pleistocene to Recent in age. As this surface is traced westward, however, it is found to be bordered on the north and west by escarpments capped with deposits of Tertiary age. It is difficult to explain this age transition of the deposits underlying such a continuous surface. In the area between the well-exposed Tertiary localities in McPherson County and the demonstrable late Pleistocene localities in southern Sedgwick County and Sumner County, the age of the deposits underlying this surface is not known with certainty.

An extensive and relatively flat plain underlain by unconsolidated deposits, believed to be in large part of Pleistocene age, occurs west of the "*Equus* beds area" in Rice, Reno, Kingman, Pratt, Stafford, Barton, and southern Ellsworth counties. This large area has not been studied in sufficient detail to differentiate Pliocene from Pleistocene strata. The outcrops are poor and are nonexistent over much of the area; therefore, detailed studies must be made with the aid of subsurface data. Bruce F. Latta is now studying the ground-water geology and subsurface stratigraphy of these deposits in Barton and Stafford counties.

The clarification of these problems in central Kansas requires not only more extensive lithologic studies and intensive collecting of fossils but also determination of the drainage history of that part of the state during Pleistocene time. As the important Pleistocene deposits, exclusive of loess, in central, north-central, and northwestern Kansas are stream deposits that seemingly accumulated along major valleys, their proper interpretation is intimately related to the drainage history of the area. That the drainage history was not simple, seems evident. The known facts have been summarized elsewhere,[46] but many aspects of this history are unknown.

NORTHWESTERN AND NORTH-CENTRAL KANSAS

The Pleistocene beds of northwestern and north-central Kansas consist largely of stream deposits on terraces and filling minor valleys, and the overlying younger loess. These strata seemingly fall into three main categories. (1) The first group of beds consists of stream deposits considered Lower Pleistocene that accumulated as fillings in broad valleys. The lithology of these beds indicates that depositing streams flowed eastward across this part of the state, carrying detritus from the Rocky Mountain region. Sediments of local origin accumulated simultaneously in tributary valleys. These deposits may in part represent outwash from glaciers in the Rocky Mountains. Poorly rounded boulders nearly 2 feet in diameter have been observed near the

[45] Personal communication from Claude W. Hibbard.

[46] Frye, Leonard, and Hibbard, ftn. 43 (1943).

Colorado line,[47] and it is difficult to explain the presence of such blocks except as outwash from an adjacent glacier to the west. Along such major valleys as the Smoky Hill, Saline, and Solomon, there are several terrace levels, the coarse clastics in each case being capped with laminated to massive fine sand and silt. (2) The second group of sediments, occurring generally on the interstream areas, comprises the heterogeneous aggregation of beds below the horizon of a prominent soil zone in north-central Kansas. These deposits range from clay to gravel and in many cases occur topographically above the level of the prominent terraces along the major valleys. Generally they are composed of rock types common to the adjacent bedrock, but in some cases rock fragments may have been transported for long distances. Meager fossil evidence indicates that these beds are younger than the most prominent high terrace of the major valleys. (3) The third group of deposits consist of a general blanket of loess—or silt—that covers the major part of northwestern Kansas and the interstream upland areas of north-central Kansas.

The Pleistocene strata of this part of Kansas have not been studied in detail. Logan[48] noted Pleistocene terrace gravels that he called "Salt Creek gravel beds" in Russell County. Robert Hay[49] described the "Plains marl" or "Tertiary marl," and N. H. Darton[50] stated that loess covered much of northwestern Kansas. Pleistocene terrace deposits have been mentioned in several State Geological Survey reports[51] dealing with this part of the state, and Wing[52] named and described the Belleville formation, which may prove to include beds of Pleistocene age, in Republic County. The widespread loess and lower sand and gravel were described and named the Sanborn formation by Elias[53] in 1931. The stratigraphy, lithology, and fauna of this formation have recently been studied in a reconnaissance manner over a large part of northern Kansas.[54] Lugn[55] objected to the use of "Sanborn" and contended that (presumably exclusive of the prominent terrace beds) "Loveland" and "Peorian" should be used for these Pleistocene deposits. The stratigraphic names in current use for the Pleistocene beds in north-central and northwestern Kansas are: "Sanborn formation" for the extensive loess and associated lower beds extending from central Jewell County westward to the Colorado line; "Belleville formation" for the beds described in Republic Coun-

[47] Elias, p. 163 of ftn. 1 (1931); Hibbard, Frye, and Leonard, "Reconnaissance of Pleistocene Deposits in North-Central Kansas," *Kan. Geol. Surv. Bull. 52*, Part I (1944), measured section No. 17.

[48] Ftn. 31 (1897).

[49] P. 101 of ftn. 30 (1893); "Preliminary Report on the Geology of Norton County, Kansas," *Kan. Acad. Sci. Trans.*, Vol. IX (1885), pp. 17–24.

[50] "The Geology and Underground Water Resources of the Central Great Plains," *U.S. Geol. Surv. Prof. Paper 32* (1905), p. 155.

[51] W. W. Rubey and N. W. Bass, "The Geology of Russell County, Kansas," *Kan. Geol. Surv. Bull. 10* (1925), p. 21; Bass, "The Geology of Ellis County, Kansas," *Kan. Geol. Surv. Bull. 11* (1926), p. 16; K. K. Landes, "Volcanic Ash Resources of Kansas," *Kan. Geol. Surv. Bull. 14* (1928), pp. 1–58; Monta E. Wing, "The Geology of Cloud and Republic Counties, Kansas," *Kan. Geol. Surv. Bull. 15* (1930), pp. 14–18; Landes, "The Geology of Mitchell and Osborne Counties, Kansas," *Kan. Geol. Surv. Bull. 16* (1930), pp. 1–55; Landes and R. P. Keroher, "Mineral Resources of Phillips County," *Kan. Geol. Surv. Bull. 41*, Part VIII (1942), pp. 277–312.

[52] Pp. 19–21 of ftn. 51 (1930).

[53] Ftn. 1 (1931).

[54] Leonard and Frye, "Additional Studies of the Sanborn Formation, Pleistocene, in Northwestern Kansas," *Amer. Jour. Sci.*, Vol. CCXLI (1943), pp. 453–62; Hibbard, Frye, and Leonard, pp. 1–28 of ftn. 47 (1944).

[55] "The Pleistocene Geology of Nebraska," *Neb. Geol. Surv. Bull. 10* (1935), p. 196.

ty by Wing; and the restricted "McPherson formation" for the high and intermediate terrace beds along Smoky Hill Valley westward to Ellis County. Hibbard[56] has applied the name "Meade formation" at two localities to terrace deposits of the Smoky Hill Valley where they contain fossil vertebrates which he has correlated with one of the faunal zones of that formation in the Meade Basin area. Formal stratigraphic names have not been applied to the terrace deposits along such valleys as the lower Saline, lower Solomon, Republican, and others.

PRE-LOESS DEPOSITS

Stream deposits, which occur stratigraphically below the loess, cover a large part of the time span of the Pleistocene. The main terrace deposits, although possibly nowhere as old as Nebraskan, nevertheless are shown by their fauna to represent early Pleistocene deposition. The intermediate terraces along the major valleys are somewhat younger and may represent several ages. At some places the strata immediately below the soil zone or the loess are shown by their contained fossils to be only slightly older than the loess, which is most certainly stratigraphically high within the Pleistocene.

The development and preservation of the major terraces were probably dependent on two prime factors: (1) the drainage patterns, subsequent stream piracies, and adjustments of the streams during the early Pleistocene and (2) the bedrock of the particular area. The drainage history of this part of Kansas is inadequately known, but the bedrock control is quite apparent. Where these valleys cross areas of relatively resistant and homogeneous bedrock, such as the Greenhorn limestone, the ancient valley walls of the broad graded valleys in which these sediments were deposited are well preserved. Where the major valleys cross areas of nonresistant bedrock, such as the Carlile and Pierre shale, these former valley walls are indistinct. Where they cross areas of Dakota bedrock, "false" valley walls are developed by resistant lenses of sandstone within this dominantly shale and clay formation, thus largely obscuring the outlines of the former valleys and allowing extreme dissection and destruction of the former valley floor deposits.

From the foregoing observations it seems that the pre-loess Pleistocene deposits of northwestern and north-central Kansas represent an aggregation of local deposits along present and former valleys of various sizes. They range in age throughout much of the Pleistocene; although many local deposits are of different ages, many probably are equivalent in age. As discussed earlier in this paper, it is uncertain which of these deposits are deserving of stratigraphic names and whether local names should be used in all cases or if names such as "Meade formation" should be used to include a variety of these local deposits whose age is judged to lie somewhere within the time span of the Meade. It is not the purpose of this paper to propose modification of present common usage of stratigraphic names for the deposits discussed.

The major terrace deposits generally have not yet yielded faunas of sufficient size and significance to permit detailed correlations within the Pleistocene. Large faunas have been collected from the high terrace deposits at three localities in Russell and Lincoln counties, and Hibbard[57] has correlated them with definite beds of the Meade formation in

[56] Pp. 714–15 of ftn. 19 (1944).

[57] Pp. 740–42 of ftn. 19 (1944).

Meade and Clark counties, Kansas. The important terraces constitute well-defined levels in many places, and it is probable that if they were studied in detail physiographically they could be traced down the major valleys to Kansas River Valley through the pre-Pleistocene divide area in the Flint Hills, thus effecting a correlation with the glacial section in northeastern Kansas.

The age of the more or less isolated and local deposits that occur on the uplands below the horizon of the important Sanborn soil zone ranges within wide limits. These deposits constitute the fillings in minor depressions and valleys between the major valleys. Elias[58] and Lugn[59] have stated that these deposits in some places are equivalent to beds called "Loveland" in Nebraska. In actuality, each exposure presents its own special problems of correlation due to the discontinuous nature of the beds. It seems that the retention of the broadly defined Sanborn formation to include these beds is necessary until more detailed studies have been made.

PROBLEMS OF THE LOESS

Light tan-gray loess—or silt—mantles much of the uplands of northwestern Kansas and constitutes the most extensive, homogeneous, and easily recognizable lithologic unit within the Pleistocene deposits of Kansas. This loess attains a maximum thickness of more than 100 feet and exceeds 50 feet in thickness over wide areas. Elias' Sanborn formation[60] includes the loess, and Lugn[61] has referred to it as equivalent to the Peorian. The stratigraphy and paleontology of the deposit have been described recently.[62] This extensive loess has been observed resting on Cretaceous formations ranging from Dakota to Pierre, on Pliocene deposits, and on a variety of older Pleistocene beds. In north-central Kansas it is separated from older beds by a well-developed soil zone, and one or more poorly developed soil zones have been observed within the loess. Generally the loess is structureless, but in some places an indistinct bedding has been observed. Concretions of calcium carbonate are abundant throughout the loess in the eastern part of the area; they gradually decrease in abundance westward and are quite rare near the Colorado line.

The problems of the loess are twofold: (1) the age and correlation of this deposit within the Upper Pleistocene and (2) the origin of the sediments. Northwestern Kansas is not immediately adjacent to the border of an Upper Pleistocene ice sheet or to a stream valley that was heavily laden with late Pleistocene outwash; it is possible that these factors—believed by many geologists to be important to the accumulation of loess elsewhere—exerted little influence on this loess deposition. It is not known, however, to what distance from the ice front the various glaciers of the Wisconsin may have influenced erosion and deposition. It seems that the history of this region during the last seventy years furnishes a clue to the manner of deposition, if not the source, of the High Plains loess generally. During the widely publicized "dust-bowl" days of a few years ago, the High Plains were subjected to a series of dust storms and dust "blows." These "dusters," prevalent from New Mexico

[58] "Geology of Rawlins and Decatur Counties with Special Reference to Water Resources," *Kan. Geol. Surv. Mineral Resources Circ.* 7 (1937), p. 7.

[59] Ftn. 55 (1935).

[60] Ftn. 1 (1931).

[61] Ftn. 55 (1935).

[62] Leonard and Frye, ftn. 54 (1943); Hibbard, Frye, and Leonard, ftn. 54 (1944).

and Texas to Nebraska and farther north, moved enormous tonnages of silt over this general area. Silt that in some places cannot be distinguished from Sanborn loess is known to have accumulated against obstruction on the High Plains surface to depths of more than 6 feet during a few years, although the maximum depth of silt accumulation on the unobstructed surface was probably much less. Occasional rains created ponds that contained water for months at a time in shallow depressions on the upland surface, in spite of general drought conditions. If periodic recurrences of such conditions existed throughout late Pleistocene and Recent time (and there is historical evidence for three such periods), a deposit such as is represented by this body of loess could be expected. It is the opinion of the writer that at least the part of the Sanborn loess that occurs above the upper discontinuous and locally indistinct soil zones is not related to glaciation and was deposited after the retreat of the last Pleistocene ice sheet. The molluscan fauna, studied by Leonard, could have existed under conditions of slow intermittent accumulation of silt, for the most part by wind action, but locally trapped in shallow upland depressions that were occasionally filled with water.

R. J. Russell[63] recently has reviewed the definition and usage of the term "loess" and the various hypotheses for its origin. As explanation of the Mississippi Valley loess, he has advanced a hypothesis of colluvial origin by "loessification" of water-deposited sediments, and he concludes that most loess deposits are not of eolian origin. C. C. Williams is now making detailed studies of terrace deposits along the Arkansas Valley south of Wichita and has generously supplied the writer with data on this area. The deposits will be discussed in detail in a forthcoming report by Williams on the ground-water resources of the lower Arkansas Valley in Kansas. These deposits, like those studied by the writer along some other valleys of central Kansas, seem to meet Russell's genetic requirements for loess. In some cases this material has been called "loess,"[64] but, because of the presence of pebbles and its probable fluvial (or fluvial-colluvial) origin, the present writer[65] has considered it a part of the Pleistocene terrace deposit.

Russell's hypothesis of origin cannot be accepted for the extensive loess—or silt—deposits that comprise the upper part of the Sanborn formation of northwestern Kansas for the following reasons. (1) This loess blankets the uplands over an area of many thousand square miles and underlies to depths of 40 feet or more the relatively flat upland surface of the High Plains; thus it cannot be the result of colluvial processes. Test holes drilled in Thomas County reveal the loess to be thickest under the flat interstream areas, and field observations show the valley slopes to be blanketed with colluvial material, mostly derived from the loess but also containing sand and pebbles from the underlying Ogallala formation. (2) Russell stated that the faunas of the Mississippi Valley loess are living in the area today and that the shells were incorporated in the material after the original deposition by stream action. This statement is not applicable to northwestern Kansas loess because, although more than 80 per cent of its molluscan fauna is now living in the

[63] "Lower Mississippi Valley Loess," *Bull. Geol. Soc. Amer.*, Vol. LV (1944), pp. 1–40.

[64] Bass, "The Geology of Cowley County, Kansas," *Kan. Geol. Surv. Bull. 12* (1929), pp. 109–10.

[65] Frye, Leonard, and Hibbard, ftn. 43 (1943).

area,[66] the remains of vertebrates (*Citellus richardsonii, Microtus pennsylvanicus, Parelephas* sp., *Platygonus leptorhinus*) that do not now live in northwestern Kansas have been collected from the material.[67]

Along major valleys, slope deposits that are in part probably the result of mass creep, as Russell suggests, give a false idea of the thickness of the loess. Elias[68] clearly excluded these deposits from the Sanborn formation in his original definition, although he did refer to them as loess.

As to the age of the loess, it has been impossible to differentiate with certainty the Recent, or upper, part from the older, more extensive part. The probable conditions of deposition previously discussed have an important bearing on age determination and stratigraphic correlation. In the upper Mississippi Valley region the Loveland loess is restricted to an interglacial interval, and the Peorian loess was probably deposited during the closing phases of the Iowan glaciation, and both loess deposits in that area occur stratigraphically between till sheets. The High Plains loess is not so restricted and probably represents an age span much greater than the Peorian, although it may include the time represented by the making of this loess sheet.

Elias[69] has suggested that in some places the red deposits below the soil zone, or the upper gray-tan loess, are eolian in origin and equivalent to the Loveland formation of Nebraska. The beds below the horizon of the prominent soil zone may include age equivalents to the Loveland of Nebraska, but over much of the area under consideration these deposits are not loess-like and are not of eolian origin.[70]

SOUTHWESTERN KANSAS

The thickest sequence of extensive nonglacial Lower Pleistocene deposits in Kansas probably occurs in the southwestern part of the state, particularly in and adjacent to the Meade Basin in Meade, Clark, Ford, Gray, Haskell, and Seward counties, Kansas, and Beaver County, Oklahoma. Until recently, the Pleistocene beds of southwestern Kansas were not clearly differentiated from the Tertiary deposits of the area. During each field season from 1936 through 1944, Claude W. Hibbard has collected fossil vertebrates from the Meade Basin region, and he has published descriptions of many Pleistocene forms from this area.[71] Smith made a reconnaissance of the geology of southwestern Kansas, and in 1940 he[72] definitely separated from the Tertiary and described as Pleistocene several formations (Odee formation, Jones Ranch beds, *Equus niobrarensis* beds, Kingsdown formation) in the Meade Basin region. In 1941 Frye and Hibbard[73] described the Pleistocene beds that occur in the Meade Basin, redefined Cragin's "Meade gravels" as the Meade formation and Cragin's "Kingsdown marl" (Smith's Kingsdown forma-

[66] Leonard, personal communication.

[67] Hibbard, personal communication.

[68] P. 180 of ftn. 1 (1931).

[69] Ftn. 58 (1937).

[70] Leonard and Frye, ftn. 54 (1943); Hibbard, Frye, and Leonard, ftn. 54 (1944).

[71] Hibbard, "Notes on Some Vertebrates from the Pleistocene of Kansas," *Kan. Acad. Sci. Trans.*, Vol. XL (1938), pp. 233–37; "Notes on Some Mammals from the Pleistocene of Kansas," *Kan. Acad. Sci. Trans.*, Vol. XLII (1939), pp. 463–79; "A New Pleistocene Fauna from Meade County, Kansas," *Kan. Acad. Sci. Trans.*, Vol. XLIII (1940), pp. 417–25; "A New *Synaptomys* from the Pleistocene," *Kan. Univ. Sci. Bull.*, Vol. XLI (1940), pp. 367–71; "The Borchers Fauna, a New Pleistocene Interglacial Fauna from Meade County, Kansas," *Kan. Geol. Surv. Bull. 38*, Part VII (1941), pp. 197–220; pp. 707–54 of ftn. 19 (1944).

[72] Smith, ftn. 2 (1940).

[73] Pp. 389–424 of ftn. 5 (1941).

tion) as the Kingsdown silt, and listed the known faunas of these formations. More recently the Pleistocene deposits of Morton, Hamilton, and Kearny counties have been described by McLaughlin;[74] of Stanton, Finney, and Gray counties by Bruce F. Latta;[75] of Ford County by H. A. Waite;[76] of Meade County by Frye;[77] and of parts of Seward, Meade, and Clark counties by Hibbard.[78]

The Pleistocene beds of southwestern Kansas consist of sediments deposited in and adjacent to several relatively large structural basins, particularly the Meade Basin; fillings of sinkholes of various types and sizes, including such extensive solutional subsidence areas[79] as the Ashland-Englewood basin in Clark and Meade counties, Kansas, and Harper and Beaver counties, Oklahoma; as channel fillings unconformably above older Pleistocene, Tertiary, or Permian beds; as terrace deposits along Arkansas, Cimarron, Medicine Lodge, and other valleys; and as eolian deposits of loess and dune sand. These Pleistocene strata contain a variety of clastic sediments that include such materials as plastic clay, volcanic ash, silt, sand, and coarse gravel. Thin-bedded redbeds have also been observed which contain selenite crystals having a maximum length of more than 10 inches. The surface of 80–90 per cent of the southwestern corner of Kansas (from Clark County to the Colorado line and from the Arkansas Valley south to Oklahoma) is underlain by deposits of Pleistocene age. Pleistocene sediments also occur at many places in other parts of southwestern Kansas.

MEADE FORMATION AND RELATED DEPOSITS

The thickest, most varied, and most broadly inclusive Pleistocene formation in southwestern Kansas is the Meade formation. Cragin[80] originally named the "Meade gravels" in 1896. The Meade formation, which presumably contains Cragin's "Meade gravels," was described by Frye and Hibbard[81] in 1941, and further data were presented by Frye[82] in 1942 and by Hibbard[83] in 1944. The Meade formation typically contains coarse gravel at the base that attains in exposures a maximum thickness of more than 35 feet. Well data indicate that this basal sand and gravel may exceed 100 feet in thickness in the subsurface. In some places the basal gravel is well cemented with calcium carbonate. The basal gravel is typically overlain by caliche, varicolored silt, clay, and fine sand, which are believed to include Smith's Odee formation. The part of the Meade that represents Smith's Odee formation ranges greatly in thickness but probably attains a maximum of more than 100 feet. Cragin's "Pearlette ash," which ranges in thickness from a featheredge to about 15 feet, and, in addition,

[74] Thad G. McLaughlin, "Geology and Ground-Water Resources of Morton County, Kansas," *Kan. Geol. Surv. Bull.* *40* (1942), pp. 80–89; "Geology and Ground-Water Resources of Hamilton and Kearny Counties, Kansas," *Kan. Geol. Surv. Bull.* *49* (1943), pp. 140–52.

[75] "Geology and Ground-Water Resources of Stanton County, Kansas," *Kan. Geol. Surv. Bull.* *37* (1941), pp. 80–86; "Geology and Ground-Water Resources of Finney and Gray Counties, Kansas," *Kan. Geol. Surv. Bull.* *55* (1944), pp. 171–83.

[76] "Geology and Ground-Water Resources of Ford County, Kansas," *Kan. Geol. Surv. Bull.* *43* (1942), pp. 162–68.

[77] "Geology and Ground-Water Resources of Meade County, Kansas," *Kan. Geol. Surv. Bull.* *45* (1942), pp. 103–12.

[78] Pp. 707–54 of ftn. 19 (1944).

[79] John C. Frye and Stuart L. Schoff, "Deep-seated Solution in the Meade Basin and Vicinity, Kansas and Oklahoma," *Amer. Geophys. Union Trans.* (1942), pp. 35–39.

[80] P. 54 of ftn. 34 (1896).

[81] Pp. 411–19 of ftn. 5 (1941).

[82] Pp. 103–9 of ftn. 77 (1942).

[83] Pp. 709–45 of ftn. 19 (1944).

30 feet or more of overlying silt, sand, and caliche constitutes the remainder of the type Meade. This sequence of beds comprises the major part of the Pleistocene strata in the Meade Basin. The deposits are persistent over several counties and have been loosely referred to as the "Lower Meade formation."

A varied assemblage of channel and sinkhole deposits of gravel, sand, silt, clay, volcanic ash, and caliche occurs above typical or "Lower" Meade beds and below the Kingsdown silt in the Meade Basin area. These deposits were included within the Meade formation by Frye and Hibbard and have been loosely referred to as the "Upper Meade." Smith's *Equus niobrarensis* beds and Jones Ranch beds (which have yielded the Jones fauna) are believed to be included in this upper part of the formation. In every exposure where "Upper" Meade deposits have been observed in contact with the "Lower" Meade, they are separated by a prominent unconformity which may represent an erosion interval as long or longer than that represented by the unconformity between the Meade and Kingsdown. These discontinuous and varied deposits that constitute the "Upper" Meade probably are not all of the same age and were included originally within the larger formation primarily as an expedient in mapping. Pleistocene beds of sand, gravel, silt, caliche, and volcanic ash, which are probably equivalent in age to part of the Meade formation, are believed to occur in all the counties of southwestern Kansas; however, they have been specifically referred to this formation only in and immediately adjacent to the Meade Basin. Deposits of equivalent age west of Meade and Seward counties have been generally referred to simply as "undifferentiated Pleistocene beds."

Two major stratigraphic problems are presented by the Meade formation: (1) the proper delineation and disposition of the beds that have been considered as the Upper Meade and (2) the accurate delineation of the Pliocene-Pleistocene boundary line in the Meade Basin.

The retention of these younger Pleistocene beds within the Meade formation seems justified only because of the inadequacy of field data and convenience in mapping. Stratigraphically they are distinct from the underlying deposits, and they have yielded a younger fauna at several localities. Subsequent studies may result in their being placed in a separate formation. The second of the major problems of the Meade is more far-reaching in its implications. The various considerations discussed earlier in this paper in regard to the establishment of the Pliocene-Pleistocene boundary are especially applicable to this area, and the special problems involved in the dating of the Rexroad beds have been discussed in some detail. A conclusive solution to this problem is of paramount importance for a proper understanding of the Cenozoic beds in western Kansas and throughout the Great Plains region.

KINGSDOWN SILT

The Kingsdown silt occurs widely over southwestern Kansas and is developed typically in southern Ford County, northern Clark County, and northern Meade County. It consists of thin-bedded fine sand and silt and massive silt. The beds are various shades of tan and light gray in color, and they mantle the uplands and cover much of the floor of the Meade artesian basin. In many places these beds seem to have been deposited by water, and in the Meade artesian basin they probably represent lake silts deposited in the late Pleistocene "Meade lake."[84] The upper massive and struc-

[84] Frye, pp. 31–32 of ftn. 77 (1942).

tureless part of the formation, generally less than 25 feet thick where it blankets the High Plains surface, has been referred to as loess and is believed to be an eolian deposit.

The Kingsdown is generally believed to be Pleistocene in age. It is mostly not fossiliferous, but it overlies the Meade formation at many places, and its topographic position on the uplands above deeply incised valleys seems to indicate that it is older than Recent. The uppermost deposits of wind-blown silt at some localities, however, were probably deposited during Recent time.

Hibbard[85] has subdivided the formation in northern Clark County into the Upper and Lower Kingsdown. The beds that he calls "Lower Kingsdown" lie unconformably below the more extensive Upper Kingsdown and are known only from northern Clark County and Comanche County. The Lower Kingsdown deposits are dominantly massive light-buff to tan silt. They differ lithologically from the beds outcropping west of Clark County which are considered "Upper" Meade, but their stratigraphic occurrence indicates that they may be equivalent in age to some of the beds that have been placed in the Meade formation. The disconformity below the Lower Kingsdown silt and above a thick bed of volcanic ash (correlated with the "Pearlette ash member" of the Meade formation) is poorly exposed and has been studied only in one small area.

GERLANE FORMATION AND ASSOCIATED BEDS

The Gerlane formation was named by Garold Louis Knight[86] from the village of that name on Medicine Lodge River, in the southeastern part of Barber County. In an unpublished manuscript in the files of the State Geological Survey of Kansas, Knight describes this formation in Barber County as follows:

> The Gerlane formation is a fluvial deposit that occurs in the valleys and on the lower slopes of the hills. The material of the formation varies in size from clay to gravel, the major part being in the finer grades of this range. A layer of coarse fragments commonly occurs at the bottom of the formation. Noticeable cementation of the fragments was seen in very few places. Erosional cuts in the finer material stand with loess-like steepness. The thickness of the Gerlane formation varies from zero at its margins to a known maximum of 104 feet. The latter thickness was revealed by core drilling on the plain south of the town of Sharon. The formation is about 60 feet thick in the Medicine Lodge River valley near Kiowa, and at least 57 feet thick in the valley of a tributary to Elm Creek at a point about 8 miles north of Medicine Lodge.
>
> Sediment derived from the erosion of the older surface rocks, chiefly the Permian and Tertiary, has been reworked and redeposited to form the Gerlane. Hence its character varies from place to place and at any particular location may resemble either the Permian or the Tertiary formations, depending upon which of them supplied the major part of the sediment. Where Permian rocks have supplied most of the materials that make up the Gerlane formation, it is red and fine-grained like the Permian, but can be distinguished from the latter by the presence of coarse sand grains derived from the Tertiary and by the fact that bedding follows the sloping surface upon which deposition occurred. Where the Gerlane is composed primarily of material from the Tertiary, the resemblance of the two formations may be quite close. As a rule, however, the Gerlane occurs at much lower elevations and has more distinct bedding and cross-bedding than does the Tertiary.

The term "Gerlane formation" has been applied to Pleistocene beds only in Barber County, Kansas. These deposits underlie the floor of the lowland surface, discussed elsewhere in this paper, that extends westward as far as eastern Meade County, topographically below the Tertiary-capped upland to the north. In Clark County and eastern Meade County, Kansas, and in Beaver and

[85] Pp. 745–49 of ftn. 19 (1944).

[86] "Gerlane Formation" (abstr.), *Geol. Soc. Amer. Proc. for 1933* (1934), p. 91.

Harper counties, Oklahoma, deposits of similar character and thickness underlie the floor of the extensive compound solutional-subsidence area that has been referred to as the Ashland-Englewood Basin. These deposits (loosely referred to as the "Ashland-Englewood terrace beds") are shown by their physiographic relations, invertebrate fossils, and a few fossil vertebrates to be late Pleistocene in age, but their exact correlation with the named formations in the Meade Basin and elsewhere in central and western Kansas is unknown. The Tertiary Ogallala formation caps the adjacent uplands north and west of the Ashland-Englewood Basin, and at a few places Kingsdown silt occurs on the uplands adjacent to the basin rim. Upper Pleistocene terrace deposits occur at an elevation higher than the basin floor along the major valleys entering this basin from the west and northwest. The age and stratigraphic and physiographic relations of the Gerlane formation and associated beds are problems deserving further study.

TERRACE DEPOSITS

Extensive terrace deposits occur along many of the major valleys of southwestern Kansas. Terraces of late Pleistocene age have been described at two levels along the Cimarron Valley,[87] and terrace deposits of supposed late Pleistocene age have been described along the Arkansas Valley.[88] Pleistocene terrace deposits that have yielded fossil vertebrates[89] occur along Salt Fork River in Barber County. These deposits and others along Medicine Lodge Valley may prove to be related in age to the Gerlane formation and associated beds. Other less extensive terrace deposits occur along many of the stream valleys in southwestern Kansas. None has as yet been studied in sufficient detail to afford a basis for exact age determination or definite correlation with more widespread Pleistocene formations. The available evidence indicates that they are all of late Pleistocene or Recent ages. The age and stratigraphic and physiographic relations of these terrace deposits present a problem deserving further study.

CONCLUSIONS

Many factual data concerning the Pleistocene stratigraphy and paleontology of central and western Kansas have been obtained during the last half-decade. These new data have been used by the several workers in this area to demonstrate the Pleistocene age of a considerable thickness and diversity of sediments formerly believed to be Tertiary in age. These recently discovered facts and new interpretations have revealed a complexity formerly unknown in the Kansas Pleistocene and have raised several hitherto unsuspected important stratigraphic problems. A review of the pertinent facts now known about the nonglacial Pleistocene deposits of central and western Kansas has been presented above. The most significant of the stratigraphic problems are summarized below.

1. Delineation of the Pliocene-Pleistocene boundary line is of regional importance. The nonmarine-nonglacial Pleistocene of Kansas is believed to be closely related to the glacial deposits to the northeast, but it has not been determined with certainty that the commonly used Pliocene-Pleistocene boundary in the Great Plains represents the same time datum as the base of the glacial Pleistocene. These deposits generally have not been correlated with the up-

87 Frye, pp. 110–12 of ftn. 77 (1942).

88 Smith, pp. 125–26 of ftn. 2 (1940); Waite, p. 168 of ftn. 76 (1942).

89 Hibbard, personal communication.

per Mississippi Valley formations, and correlations have not been attempted in this paper because of the rarity of vertebrate faunas in the upper Mississippi Valley Pleistocene beds, the discontinuous nature of all but the terrace and eolian deposits among the nonglacial formations herein discussed, the absence of adjacent glacial deposits younger than the Kansan, and the lack of detailed data in the area in Kansas peripheral to the glacial deposits.

2. In central Kansas the relationship and correlation of minor terrace deposits to the more extensive McPherson formation is not clear.

3. The several Pleistocene terraces along Smoky Hill Valley are believed to constitute a westward extension of the major abandoned valley filling, the McPherson formation, but no one terrace has as yet been correlated with the type locality.

4. The extensive upland deposits extending from northern McPherson County to the Oklahoma line are known to be of Pliocene age in northern McPherson County and of Pleistocene age in southern Sedgwick and Sumner counties, but the age and physiographic history of these seeming continuous deposits in the intermediate area are obscure. Also, the beds constituting the "Upper" Emma Creek formation in this general area, although believed to be of Pliocene age, may be partly or wholly of early Pleistocene age.

5. The Belleville formation in north-central Kansas has been assigned to the Tertiary, but some data indicate that it may be Pleistocene in age.

6. The several beds of the Sanborn formation of northwestern and north-central Kansas have not as yet been correlated with the major terraces of that region, with the McPherson formation of central Kansas, with formations of the Meade Basin, or with formations of southern Nebraska.

7. Physiographic processes operating during Sanborn time resulted in the deposition of sediments ranging from eolian loess to boulder beds. Lack of knowledge of the origin of these deposits inhibits correlations of Lower Sanborn beds throughout northwestern Kansas.

8. Determination of the age and correlation of the Rexroad member in the Meade Basin represents a major problem in southwestern Kansas.

9. Many deposits loosely referred to as "Upper Meade," and also some isolated sinkhole fillings, have not been correlated with one another or with other formations of western Kansas.

10. The correlation of the Lower Kingsdown silt is a problem deserving continued study. It may be equivalent to or even older than some beds referred to as the "Upper Meade" formation. The Lower Kingsdown silt has been studied only in northern Clark County. The disconformity below the Lower Kingsdown silt and above a thick bed of volcanic ash (correlated with the "Pearlette ash member" of the Meade formation) is poorly exposed and has been studied in one small area only.

11. The Gerlane formation presents a problem of origin, correlation, and age within the Pleistocene.

12. Terrace deposits occur extensively in southwestern and south-central Kansas. They present a multiplicity of problems of age and correlation.

ACKNOWLEDGMENTS.—Thanks are expressed to colleagues, of the state and federal geological surveys in Kansas, who have supplied data and offered criticisms and advice. The manuscript has been read by A. C. Trowbridge, R. C. Moore, Elmer S. Riggs, Claude W. Hibbard, and A. Byron Leonard.

DISCUSSION ON THE VISCOSITY OF LAVA

CHESTER K. WENTWORTH
Honolulu Board of Water Supply, Honolulu

MAX H. CARSON
United States Geological Survey, Honolulu

AND

RUY H. FINCH
National Park Service, Kilauea, Hawaii

ABSTRACT

In 1939 attention was called by Nichols to the failure of earlier workers to apply the Reynolds number criterion and distinguish between laminar and turbulent flow in estimating the viscosity of lava from observed flow velocities. The present writers follow Nichols in his primary contention but take exception to some of his argument and much of his application of field observations in estimating viscosity of lava. It is the conclusion of the paper that even with the use of the Reynolds number and the Stanton diagram in outlining flow conditions, according to the best available data, the complexities and uncertainties of the flow of lava under field conditions preclude definitions of condition or statements of results with sufficient precision to justify presentation as numerical coefficients of viscosity in the ordinary sense.

INTRODUCTION

Because of practical difficulties in making decisive measurements of the viscosity of lava under field conditions, geologists have had recourse to various rough estimates and comparisons. Added to the weakness inherent in extrapolation from rough measurements, the lack of common experience among various geologists has further retarded attainment of any generally satisfactory values. Nichols has pointed out the error of estimates made by Becker and by Palmer, owing to their failure to recognize the necessity imposed by the Reynolds number and friction factor that the flow of lava be laminar.[1] While this position is conceded to be a valid one, certain conclusions put forward by Nichols do not seem justified.[2] Further, it is believed that the skepticism felt by many geologists toward the high viscosity ratios and their difficulty in accepting such values as real justifies a more complete step-by-step discussion of the problem.

The problem divides itself into two parts: first, what is the viscosity of lava and how precisely can we consider it has been determined for any definable condition and, second, what are the facts, methods, and validity of cross-comparison between a calculated velocity of water in a given channel and the observed velocity of lava in a comparable

[1] Robert L. Nichols, "Viscosity of Lava," *Jour. Geol.*, Vol. XLVII (1939), pp. 290–302; G. F. Becker, "Some Queries on Rock Differentiation," *Amer. Jour. Sci.*, Vol. III (4th ser., 1897), p. 29; H. S. Palmer, *Hawaiian Volcano Observatory, Monthly Bulletin*, Vol. XV (1927), pp. 1–4.

This paper has been prepared with the approval of Dr. Palmer, who indicated his willingness to leave the discussion to us and furnished useful supplementary material.

[2] The writers recognize that both Becker and Palmer failed to distinguish between turbulent and laminar flow but find no justification for Nichols' statement that they assumed the flows to be turbulent. Since they both assumed a reciprocal relationship between the first power of velocity and viscosity, which is the chief feature of laminar flow, it would seem more correct to say that they tacitly assumed all the flows to be laminar, though Palmer used the Chezy formula for calculation of velocity of water.

channel as means of estimating the viscosity of the lava. It is proposed to examine the latter of these problems first.

The flow of liquids, and especially water, in channels and in pipes has been the subject of study from the time of Archimedes. Because of the preponderant study of water in channel sizes and grades that were of practical importance, the formulas earlier developed were of practical, empirical sorts having various natural constants, such as gravity, density, and viscosity, lumped in one or more composite constants. Most of the flow of early, everyday importance was of the turbulent type. Discrimination between turbulent flow and laminar flow came gradually. Through the experimental work of Hagen and Poiseuille about 1840 and the subsequent mathematical formulation of the nature of laminar flow in tubes, this type of flow became perhaps the best understood and most rationally formulated of any phenomenon in the field of fluid mechanics.[3]

Study of turbulent flow in channels has been carried on over a long period, with development of various formulas, many with some rational framework, but all with ultimate reliance on certain empirical constants.[4] The work of Osborne Reynolds, published in 1883, is fundamental to an understanding of the relationship between laminar flow and turbulent flow.[5] It is very puzzling to any scientist when he realizes how long a period is commonly required before an important and widely applicable discovery or formulation is generally in use among students or workers in a given field. Despite the fairly prompt recognition of the basic character of Reynolds' work by specialists soon after 1883 and the universal applicability of the Reynolds number in defining the so-called "critical zone" and the discrimination between laminar and turbulent flow, there is very little mention of it in American textbooks of hydraulics prior to 1925, and it is still not generally understood.[6]

[3] Ralph W. Powell, *Mechanics of Liquids* (New York: Macmillan Co., 1940), p. 167.

[4] "Hydraulics," *Encyclopaedia Britannica* (11th ed., 1910), XIV, pp. 68–84.

[5] *Phil. Trans. Roy. Soc. London* (1883).

THE REYNOLDS NUMBER

Before attempting to indicate the importance of the Reynolds number, some simple definitions must be offered. Laminar flow, for our present purpose, is best defined as flow in which the resistance to motion is due to viscous, shearlike deformation. The impelling force is a pressure or potential difference in which the density or unit mass of the liquid is involved, but the density is not involved in the resistance to motion. Turbulent motion, on the other hand, is flow in which the resistance to motion is that due to the inertia of the fluid and hence is dependent on density or unit mass. In turbulent flow the viscosity is a negligible factor and is only significant in determining its lower limit. In any given instance laminar flow takes place at lower velocities than turbulent flow. Both types of flow may exist in the same channel, with turbulent flow coming in at points where the channel is rough or

[6] "For many years the far-reaching implications of Reynolds' work were ignored by practical engineers, most of whom were entirely ignorant of it. Both mechanical and civil engineers continued to estimate the friction loss in pipes by methods that had been in use since Chezy. But the appearance of the new branches of chemical engineering, petroleum engineering, and aeronautical engineering, created a demand for methods which would apply to [fluids] other than water. The old-style hydraulics men had apparently never considered whether or not the [formula for friction loss] applied to fluids other than water. . . . . It was only after some 30 or 40 years that Reynolds' work began to be studied by practical men" (Powell, p. 173 of ftn. 3 [1940]).

changes direction. Relative widths of the zones of laminar and of turbulent flow depend on the magnitudes of the velocities, densities, and viscosities as they enter into the Reynolds number.

For laminar flow we have the Jeffreys formula quoted by Nichols,[7]

$$V = \frac{g \sin A d^2 p}{3u},$$

from which we see that velocity is proportional to the acceleration of gravity, to the slope ($\sin A$), to the square of the depth or hydraulic radius, and to the density of the liquid, and inversely proportional to the viscosity (u). On the other hand, for turbulent flow, we have the very widely used Chezy formula,

$$V = c\sqrt{rs},$$

where we find velocity proportional to the square root of both the hydraulic radius and the slope.

Now we should recognize, in this laymen's discussion, that while in laminar flow we have a form of fluid movement in which viscosity is the ruling factor and that in turbulent flow it is the inertia of the mass of the liquid which is the ruling factor, we cannot say that either of these factors becomes zero in the realm of the other. Actually, it is only that the effect of one becomes negligible in the range where the other is preponderant.

It might occur to us that inertia would tend to become preponderant over viscosity with increase of density, velocity, and diameter or depth of channel and with decrease of viscosity. A formula showing such a tendency could be written as

$$R = \frac{pVd}{u},$$

where p is the density, V is velocity, d is diameter, and u is viscosity. The value R is the very important Reynolds number, which is, of course, based on more rigorous mathematical reasoning than is given here and which has been checked by the so-called "dimensional analysis" used in problems in similitude and found to be nondimensional. This means that it is independent of any particular values of units of length, mass, and the like and hence is of very general utility. It is applicable to various other physical phenomena where viscosities and masses of liquids are involved, one example being in the problem of the falling of spheres through liquids, where it serves to discriminate between the low velocity fall resisted by viscous drag (Stoke's law) and the higher velocity fall resisted by inertia (sixth-power law, for transport by streams).

It should now be explained that there is found by experiment a critical value of the Reynolds number, determined by Reynolds and many others, especially Schiller, below which the flow is laminar, but at which signs of turbulence are seen.[8] This is the point where the magnitude of the resistance due to inertia (friction of the hydraulicians) becomes comparable to the resistance due to viscosity and above which the inertia effect is predominant. It is found that, while turbulent flow tends to commence at a value for the Reynolds number of 2320, laminar flow may persist in a smooth tube and under conditions of minimum disturbance to much higher values. The limit 2320, known as the lower critical value of the Reynolds number, appears quite definite; and turbulent flow, however much encouraged by disturbance, does not persist below this value.

[7] This is basically a friction factor.

[8] R. A. Dodge and Milton J. Thompson, *Fluid Mechanics* (New York, McGraw-Hill Book Co., 1937), pp. 179–86; Powell, pp. 168–73 of ftn. 3 (1940).

The Reynolds number, then, is a kind of score, depending on the quantities' density, velocity, diameter of cross-section, and viscosity, which will tell us, for liquids of all densities and viscosities,

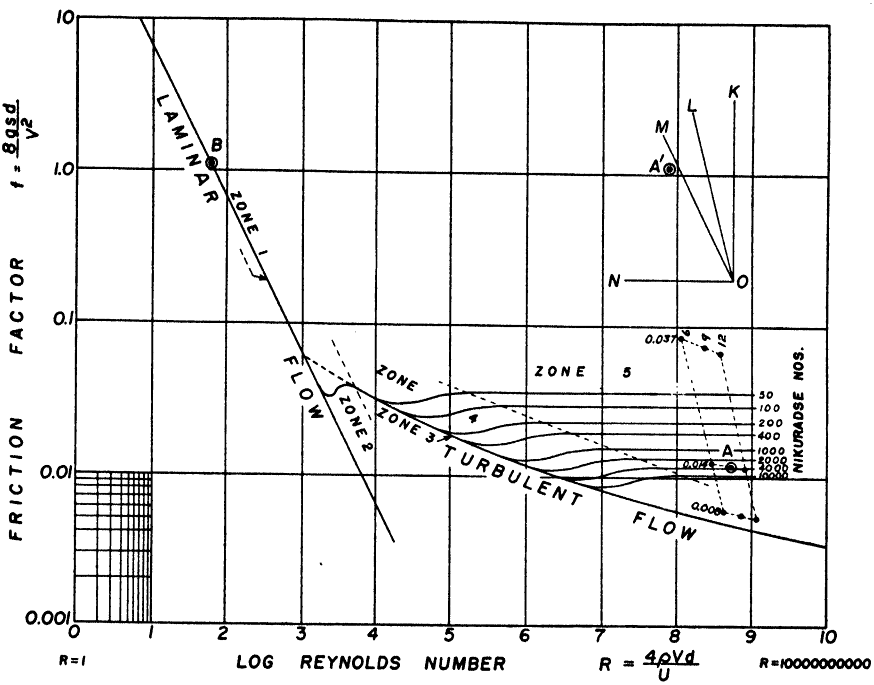


FIG. 1.—The Stanton diagram, showing the line of laminar flow, the line of turbulent flow in smooth pipes or channels, the transition zones (2 and 4), and the band of turbulent flow in rough pipes or channels. The Nikuradse numbers are the ratios of pipe diameter to diameter of roughness particles; hence they are, in reality, smoothness numbers. According to the formulas for f and R, and with channel dimensions, density and gravity fixed, migrations in various directions on the diagram have the following meanings: line OK—concurrent decrease of velocity and viscosity in same proportion, which can only mean increase in f and hence in roughness; line OL—decrease in velocity with constant viscosity; line OM—velocity decrease and viscosity increase in same proportion (this is direction of laminar flow line); line ON—velocity constant with increase of viscosity. Such migrations can have physical significance only along the lines and within the zones shown on the diagram.

and hence of a wide range of velocities and tube sizes, whether the conjoint tendency in the direction of greater velocity has a sufficient magnitude to produce, or permit, turbulent flow, or remains below the lower critical value and hence indicates laminar flow.

THE STANTON DIAGRAM

The Reynolds number is most strikingly used in connection with the friction factor, as in Figure 1. This is known as Stanton's diagram from the investigator who first used this form of plotting.[9] The abscissa scale consists of Reynolds

[9] Dodge and Thompson, p. 204 of ftn. 8 (1937). Nichols attempted to follow this form, but the full usefulness of his diagram is impaired because he did not use the more customary double logarithmic scales. This diagram involves such a wide range of numbers that Nichols' use of arithmetic scales over

numbers, $R = \frac{4pVd}{u}$, and the ordinate scale carries values of f, the friction factor from Darcy's formula, which is written

$$f = \frac{8gsd}{V^2}.$$

The Reynolds-number formula becomes multiplied by 4 when d is the hydraulic radius of a channel and used in place of D, the diameter of a pipe.[10] On this diagram it is possible to show graphically the relationships between f and R which hold for the laminar, transitional, and turbulent states of flow.[11] From the Hagen-Poiseuille law for laminar flow through tubes, it holds that in laminar flow,

$$f = \frac{64}{R},$$

a relationship which appears as a straight line in Figure 1. From experiment it has been found by various investigators that in well-established turbulent flow at higher velocities in a pipe of given roughness, the value of f is a constant, independent of velocities and of Reynolds numbers. This condition is expressed in the series of horizontal lines labeled for pipe roughness in the notation of Nikuradse. (These numbers are the ratios of the thickness of the peripheral laminar zone to the heights of the average roughness elements; hence they are really smoothness numbers, the smaller numbers applying to the rougher pipes.) It will be noted that, the smoother the pipe, the higher the Reynolds number must be before this phase is reached. The line from which the various turbulent flow lines break off has been determined to have the formula approximately as follows:

$$\log_{10} R = \frac{1}{2\sqrt{f}} - \log_{10} \sqrt{f} + 0.40,$$

and it is shown on Figure 1.[12] It should be noted that much of this discussion is based on flow in pipes. This is because the larger part of such exhaustive experimentation has been done on pipes. Because of the generalized form of plotting, the principles are applicable to the problem of flow in channels. It should be noted that the various roughness lines become horizontal at values of the Reynolds numbers that increase with decreasing roughness and that at the smoothness value of 100 the change comes at the Reynolds number of about $10^{5.5}$.

Now to summarize the characteristics of the Stanton diagram, it is pointed out by Powell that on it are shown

five distinct zones, depending on the value of the Reynolds Number. (1) The Laminar Zone, where R is less than 2,000, $f = \frac{64}{R}$, and the loss of head varies as the first power of the velocity. (2) The Critical Zone, where R is between 2,000 and say 3,000. Here the flow may be laminar but is more likely to be turbulent. . . . . (3) The Smooth Pipe Zone where the flow is turbulent but f depends only on the Reynolds Number and not at all on the roughness of the pipe wall. (4) A Transition Zone where f depends mostly on the relative roughness, but still somewhat on the Reynolds Number. (5) The Rough Pipe Zone, where f depends entirely on the relative roughness and not at all on the Reynolds Number.[13]

In so far as flow of water in pipes is concerned, it appears that all possible

a wholly inadequate range not only resulted in an awkward hyperbola in place of the usual straight laminar flow line but also achieved a grossly misleading allocation of lines B, B^1, and B^2 in his Fig. 1 (*Jour. Geol.*, Vol. XLVII [1939], p. 293).

[10] Powell, p. 202 of ftn. 3 (1940).

[11] Dodge and Thompson, pp. 202–6 of ftn. 8 (1937).

[12] Powell, p. 175 of ftn. 3 (1940).

[13] *Ibid.*, pp. 185–86.

combinations of the parameters that enter into the value f and the Reynolds number will fall within the range of the laminar flow line on the left, and thence on the smooth pipe line and the fairly broad but not indefinitely wide zone of rough-pipe values at the right. Any other allocation of flow of water in pipes is held to be impossible.

The notation of the Stanton diagram is perfectly general; it is supposed to hold for other liquids than water; and it presumably holds in all essential particulars for channels as well as pipes if comparable notation is used. But it should be recognized that no experiments have been made with water for Reynolds numbers higher than 10^7, nor with channels so large or slopes so steep as those involved in the Alika flow. Hence the data with which we are here concerned yield a Reynolds number two logarithmic cycles, or about a hundred times, higher than the field which has been fairly well explored.

From the division of the Stanton diagram into five roughly vertical zones, we conclude that comparisons of flow conditions will be more fully justified and valid when they lie in the same zone and will at least be open to more question when they involve conditions lying in two different zones. This principle is identical with the stipulation that for valid comparisons between real conditions and model-scale experiments that model experiments should be so designed that they provide similar Reynolds numbers to those of the field conditions.

COMPARISON OF FLOW CONDITIONS

We note from the location of the two points calculated for water and lava, A and B, on the Stanton diagram that they are widely separated horizontally and lie in zones 5 and 1, listed above.

Accordingly, we should proceed with considerable caution and may view even the most carefully reasoned conclusions with some skepticism. Starting with the conditions shown by the point A, we may show graphically what effect is exerted on the diagram by variations in the assumed conditions. The line 0.037–0.008 shows the effect of variation in roughness; the line 6–12 shows the effect of differences in hydraulic radius.

No one knows accurately what these values really were. An assumption was made by Palmer, based on personal acquaintance with the behavior of Hawaiian lava flows and appearance of lava slopes, that the value of 0.014 was the most probable value of n. Nichols, whose arithmetic version of the Stanton diagram is necessarily somewhat misleading, especially in the insupportable location of horizontal roughness lines at Reynolds numbers as low as 2×10^3, sees fit to conclude that "a roughness coefficient (n) of 0.037, ... would seem to be much nearer the truth," but he offers no reason for such conclusion. The present writers, as well as several other persons consulted, all familiar with the Hawaiian terrain, with the behavior of lava flows here, and with hydraulic formulas, are not inclined to follow Nichols in this conclusion and believe that Palmer's assumption of 0.014 is equally likely to be too high as too low, since the capacity of a lava flow to adjust its channel bottom in the direction of smoothness is rather great. Values of n are empirically determined: for a given shape, the roughness n, determined by experiment, might be quite different for lava and for water. We should not be misled by the fact of roughness of much of the final lava terrain, which is quite another matter.

In making a translation from point A (Fig. 1) to point B, fixed by the values of

f and *R*, it is evident from the structure of these numbers that three component parameters are permissibly variable. These are: velocity in the ratio of about 152/16.2, density in the ratio of 1/1.4, and viscosity in a ratio which is sought. The other parameters are assumed to be fixed, such as gravity, slope, and the dimensions of the channel. Since density appears only in the Reynolds number, we may at once make the change to density 1.4 by moving the point *A* to the right by that amount. If now we reduce velocity in the ratio of 152/16.2, the point *A* is to be moved to the left in the amount of about 9.4, to the position $R = 7.6 \times 10^7$. But such a reduction of velocity increases the value of *f* by the ratio 88.3, which would lead to the point A' at $f = 1.33$. This point is untenable and represents an impossible flow condition according to the Stanton diagram. No such value of the friction factor, *f*, is possible except on the laminar flow line.

The only point on the laminar flow line which satisfies the friction factor value of 1.33 is at Reynolds number $= \frac{64}{1.33}$, or $R = 4.81$. Change of the Reynolds number from point A' to point *B*, values 7.6×10^7 and 4.81, without further change of velocity, density, or diameter, can only be accomplished by an increase in the viscosity in the ratio of $(7.6/4.81) \times 10^6$. This gives approximately a viscosity in poises of 1.58×10^4. This is not offered as another measured value but rather by way of carrying to completion the reasoning started by Nichols. The value offered by Nichols of 4.3×10^4 is not calculated from the Alika observations by comparison with water but was computed directly by use of the Jeffreys formula for laminar flow, after using the reasoning of the Stanton diagram to identify the flow of lava as laminar in character. Though Nichols stated this in his paper, the trend of thought in the paper is likely to lead many readers to suppose that value to have been calculated from the Alika flow.

The question may still be asked: Why cannot the reduction in velocity take place concurrently with increase in viscosity of similar rate? We can only reiterate that such changes would lead to values of *f* and *R* which would locate a point in the portion of the Stanton diagram where no real condition of flow exists. For example, inspection of the scale of smoothness values shown by the horizontal lines of Figure 1 shows that the ratio of diameter of roughness particle to diameter of pipe of 1:1 is reached in the vicinity of $f = 0.1$. Any greater roughness than this is logically unthinkable; moreover, it is evident that, as the diameters of roughness particles approach the diameter of the pipe or channel, the remaining openings, despite their irregularity, will be so reduced in size that any remaining flow must necessarily become laminar.

Or, even if we take the indications shown by the calculated 0.008–0.037 line, we recognize that the roughness on any scale must reach limiting values long before any value of *f* which satisfies the present requirements can be reached on the Stanton diagram. Or, to take another line of reasoning, the decrease of velocity in proportion to the increase of viscosity is diagnostic of laminar flow. This change takes place along the line of laminar flow shown to the left and cannot, according to the Stanton diagram, take place at higher Reynolds numbers.

In short, despite the fact that the relations of the Stanton diagram have not been verified through all possible values of all the parameters involved, it seems most probable that the basic outlines

are valid and that flow conditions falling outside the lines and paths indicated in the five named zones are in fact nonexistent. Thus the contention of Nichols regarding the laminar flow of lava within this range of values of slope and velocity seems fully supported.

How closely, with this reasoning, may we fix the viscosity of the lava of the Alika flow? Certainly no closer than the estimate we can make of the behavior of water with the known viscosity. The variation of possible locations for point A, with n ranging from 0.008 to 0.037, corresponds to a ratio range of final values for viscosity of nearly 4; the effect of the range of hydraulic radius assumptions from 6 to 12 feet is somewhat less, but, combined by extremes with the values of n, gives a total possible range of about 10 times. This is on the assumption of strict utility of the Chezy formula and the Kutter coefficients at such slopes and velocities and depths. Even without a major error of reasoning, prudence suggests that any intermediate value chosen might be as much as two or three times, or as little as one-half or one-third the correct value. In the notation of $n \times 10^m$, this suggests that m can be indicated to the nearest unit and that n should be given a tentative integer value, but it scarcely justifies a second significant figure in n.

Similar reasoning applies to the calculation of viscosity by the Jeffreys formula from the observed velocity and the estimated depth. If Jaggar's estimates of depth as between 15 and 30 feet are accepted, the ratio 1:2 becomes squared in the formula and leads to a possible four-times variation in the result for viscosity. Moreover, the Jeffreys formula is correct for a channel of indefinite width and no drag on the sides or upper surface. It seems to us that in the case of lava flows the retarding effect of the partly congealed top surface, with its load of partly submerged blocks, may well be considerable and comparable to a considerable increase of wetted perimeter and decrease of hydraulic radius. We cannot even suggest the magnitude of this effect, but it seems that the uncertainties due to this cause might at least equal a 1:2 ratio, carrying the total plausible uncertainty here also up to eight times. If the 4.3×10^4, as calculated by Nichols from the Alika flow, is subject to such error, it might well be as low as 1.5×10^4 or as high as 1.1×10^5. In such event the second place in n is useful only for convenience in illustration and calculation.

MULTIPLE CONDITIONS OF FLOW

Finch has recently set forth observations suggesting the probable existence in some lava streams of liquid of two different sorts and probably a marked difference in viscosity.[14] Even without a sharp distinction in types, we should expect a large increase in viscosity toward the walls and bottom of any stream subject to cooling effects. If the flow is laminar, the likelihood of mixing from zone to zone is relatively small, and the probability that we may be dealing with a systematically nonhomogeneous liquid is to be faced.

We have neither field data nor the requisite background of theory for a rigorous discussion of what might be the consequences of a systematic increase of viscosity toward the walls of a conduit, but certain qualitative suggestions are possible. We may at once concede the possibility that the central part of the flow might be turbulent while a thick, peripheral part might be laminar. Effect

[14] R. H. Finch, *Volcano Letter No. 480* (1943), pp. 1-3.

of such a supposition would be that the total flow condition would fall in widened zones 2 and 3, of the Stanton diagram—hence not in the pure laminar-flow condition. Roughness would be of relatively little effect; the observed velocity of the lava flow at the surface would be too high to be diagnostic of the behavior of the laminar, under portion of the flow; and, given the surface velocity, the estimated channel size and particularly hydraulic radius might differ greatly from true values. The effect of this consideration would be to change materially the calculated velocity and the Reynolds number of water flowing in a comparable channel. Hence, any comparison with the surface lava we might make would involve shorter migrations across the Stanton diagram and the indicated viscosity would be less than before. Whether the more viscous, bottom layer would be more viscous than the results of earlier calculation is beside the point, since we would have no basis for guessing its velocity or other dimensions. If the bottom layer is indeed as viscous as the calculations seem to suggest, the possibility that the upper and exposed layers are much less viscous in itself may help to reconcile the discrepancy between the high viscosity values and the relatively high apparent fluidity which impresses all observers.

Recognition of the probable occurrence of zones of differing viscosity, while suggesting the possibility that the indicated viscosity ratio would be somewhat reduced, is of most importance in bringing realization that we may be dealing not with one liquid but with several and that we lack the resources in data and in theory to make a comparison that would be valid or to define what we desire in an answer, since the viscosity of the lava would not be a unit value.

ERRORS IN ILL-DEFINED CONDITIONS

If we have such variations in the calculation of fairly well-indicated channel condition, we must expect much more error when the field data for the 1887 flow are dealt with as loosely as they were by Nichols.[15] First of all, the acceptance of Stearns's statement that "it must have had an average velocity of 0.7 mile an hour" as a statement of rate of channel or steady flow is a definite error, since Stearns himself, in the next line after the not too unequivocal statement quoted, says: "It must have moved considerably faster in the feeding channel."[16] Use of the 0.7 mile per hour, which is obviously a rate of advance of the lava front, serves only to repudiate completely any estimate based on it. It is the less justifiable, since Nichols himself elsewhere recognizes the difference between the two rates[17] but does not, it appears to us, make sufficient amends for the misleading effect of juxtaposing the values of 4.3 and 4.77 in his table, when it is evident that the former may deviate substantially from 4 and that the latter probably does not justify tabulation in numerical form. It does not seem to us that the close similarity between the values 4.3 and 4.77 can be of the slightest real significance.

The writers do not feel any special qualification for discussing the field data in the case of the McCartys flow, to which Nichols applies some of the conclusions already discussed. But we do find it impossible to conceive the move-

[15] Pp. 295–96 of ftn. 1 (1939).

[16] H. T. Stearns and W. O. Clark, *U.S. Geol. Surv. Water Supply Paper 616* (1930), p. 74.

[17] P. 298 of ftn. 1 (1939).

ment of a lava stream 1,500 feet wide, of any thickness, whether in one unit or several, in such fashion that the flow can be considered as a continuous stream which over such width is amenable to a formula for laminar or any other type of flow. The concept of flow units is even more fundamental than Nichols appears to recognize and should certainly demand that in any 1,500-foot width there are probably several units horizontally as well as vertically, and that only a serial sectioning that it is not given geologists to achieve would deduce the order of succession of emplacement of these. In course of all this emplacement, lava would flow in certain channels, probably far narrower than the 1,500 feet, and under conditions for perhaps limited periods, which would permit determination of viscosity if we had the data. So far as our observations go on basaltic lava flows and on flow formations, it does not seem to us that the over-all dimensions offered by Nichols have any bearing whatever on the flow behavior in any stream section which would permit a viscosity determination. Hence, it appears to us that the conditions for the McCartys flow are not sufficiently defined to permit a conclusion that can be relied on—perhaps not closer than one-tenth or ten times the true value. So far as the rate of expulsion of certain Hawaiian flows is concerned, the larger error is probably the thickness factor in the volume estimate, since observers in many instances have kept fairly continuous vigil at or near the vent.

In the course of elaborating this discussion, various observations on the behavior of lava have been suggested as possible means for indicating viscosity. These include the observed splashing of lava, the breaking and spilling of bombs, the formation of Pele's hair, Pele's tears, and the like. Given sufficient opportunity for planned measurements, the observation of ripples, surges in lava lakes, the turning and foundering of floating blocks and many other behaviors could possibly be used to deduce approximate values for the existing viscosity. But in the present lack of definitive observations on any of these behaviors and the difficulties of handling, abundantly indicated by this problem already too long discussed, it is concluded that, until planned new observations can be made, any such approach would be futile.

CONCLUSION

It is the conclusion of the writers that the much-desired determination of the viscosity of basaltic lava under juvenile eruptive conditions has not been achieved with any precision having practical utility. We believe that the Nichols conclusion that the flow of the lava under the field conditions postulated was laminar is a reasonable one, and correspondingly it appears that the values for viscosity deduced by Palmer and Becker are too low. Further, it appears that various observers' impressions as to the fluidity of lava—that it "flowed like water," and such—are by no means reliable. It is probable that the ordinary human kinematic sense, unaided by instruments, is quite unable to resolve the factors involved in the Reynolds number or Stanton's diagram. Probably we shall have to discount a great part of the feeling that a viscosity value is too high or too low.

Apart from the nonhomogeneity of lava due to presence of gas, to involved blocks and crusts, and also to cooling and marked increase of "viscosity" or resistance to flow near the margins, we

should recognize also that throughout the temperature ranges shown by lavas at the surface, the processes of crystallization and mineral growth must seriously impair the comparison of lava with a true liquid. Not only is the lowering of temperature discontinuously related to heat loss, owing to heat transfer involved in the formation of successive minerals, but with the formation of those minerals the remaining fluid becomes successively a different substance. The sum total is a very complex suspension of increasing amounts of minerals in a changing residual fluid which no doubt becomes increasingly resistant to flow with lowering of temperature. It seems very doubtful if under any natural conditions of flow and of cooling in the form of a river, the mobile lava has been or perhaps can be so defined, as to its temperature and constitution from place to place, as to justify deducing its unit resistance to movement in the form of a coefficient of viscosity.

The reasoning presented in this paper, in the absence of sorely needed field data closer to the problem, supports the view that the viscosity of lava may range between 10^4 and 10^7 times that of water, or the order of from 10^2 to 10^5 poises. Data from other sources, such as laboratory determinations on dry melts at the lower temperatures, indicate viscosities that approach the high orders but start a hundred times lower at temperatures of 1,400° C. In general, we should expect gas-loaded lava flows to be more fluid than the same basalt in dry melts. It is not clear to the present writers that we can go further than to say that the viscosity of basalt in active lava-channel streams appears to fall in the range from 10^2 to 10^5 poises. In view of large uncertainties in the calculations and in the conclusions, it is not clear that such viscosity statements are of greater value than the raw data on the observed rates of flow of various lava streams, and it is believed that statement in the form of a coefficient of viscosity is unfortunately very misleading at this stage.

POST-WASATCH TERTIARY FORMATIONS IN SOUTHWESTERN UTAH[1]

HERBERT E. GREGORY
United States Geological Survey, Washington

ABSTRACT

In southwestern Utah widely exposed conglomerates, pyroclastics, and lacustrine and fluvial sediments overlie the Eocene limestones that constitute the bulk of the Wasatch formation. These consolidated and partly consolidated rocks have been grouped as the Brian Head, the Parunuweap, and the Sevier River formations.

INTRODUCTION

In southwestern Utah the consolidated and partly consolidated beds of pyroclastics, conglomerates, alluvial sands, and lacustrine silts that postdate the typical Wasatch limestones are displayed on canyon walls, in broad valleys, and on drainage divides and differ in origin, age, and mode of deposition. In reports submitted for publication by the United States Geological Survey[2] these miscellaneous deposits have been mapped as the components of three formations: Brian Head, Parunuweap, and Sevier River.

BRIAN HEAD FORMATION

DISTRIBUTION AND CHARACTER

Generally along the western edge of the Aquarius, Sevier, and Markagunt plateaus, where beds of Tertiary age are exposed in the upthrow blocks of master faults, the characteristically pink, compact limestones of the Wasatch formation are overlain by conspicuous white, regularly stratified calcareous and siliceous beds and they in turn by gray, coarse-grained, igneous conglomerates and breccias, which in places extend upward to sheets of black lava—the cap rock of the plateaus. On the southern parts of the Aquarius, Paunsaugunt, and Markagunt Plateaus the white beds are essentially limestone that include here and there varying amounts of granular and amorphous quartz, chalcedony, and decomposed lavas. On the northern Aquarius, the southern Sevier, and central Markagunt plateaus, beds in the same stratigraphic position consist chiefly of pyroclastics and highly siliceous limestones and chalcedony. Thus in a widely spread, continuous series of strata, the quartzose materials and volcanic debris increase from south to north. However, the northward change from dominant limestone to dominant tuff and ash is not regularly progressive. In places thin sheets and lenses of calcareous silts are interbedded with the thick tuffs, and in other places chunks of chalcedony and the disintegrated basic and acidic igneous rocks are scattered through thin-bedded calcareous shales.

C. E. Dutton[3] and his co-workers in the Powell and Wheeler surveys classed the stratified, white, dominantly calcareous strata that overlie the massive pink

[1] Published by permission of the director, United States Geological Survey.

[2] H. E. Gregory, "The Zion Park Region, the Paunsaugunt Region, and Eastern Iron County," U.S. Geol. Surv. papers submitted for publication.

[3] "Report of the Geology of the High Plateaus of Utah." *U.S. Geol. Surv. Rocky Mts. Region* (1880), pp. 73–74, 158–59, 199, 237–38.

limestones on the Paunsaugunt and Markagunt plateaus as "lacustrine limestones"—"the upper white limestones and marls [at the] summit of the Bitter Creek group" (Substantially the Wasatch formation)—and described the beds in corresponding stratigraphic position at the head of Parowan Canyon, in Bear Valley, and along the South Fork of the Sevier as records of volcanic activity: "fine grained marls and sandstones which have been derived from the decay of ancient lavas." In most later reports on the geology of the southern High Plateaus scant attention is given to the Tertiary strata described by Dutton; the disintegrated "ancient lavas" are barely mentioned; and the "upper white limestones and marls" are incidentally classed as "white Wasatch" and treated as a phase of deposition during undefined Wasatch time.

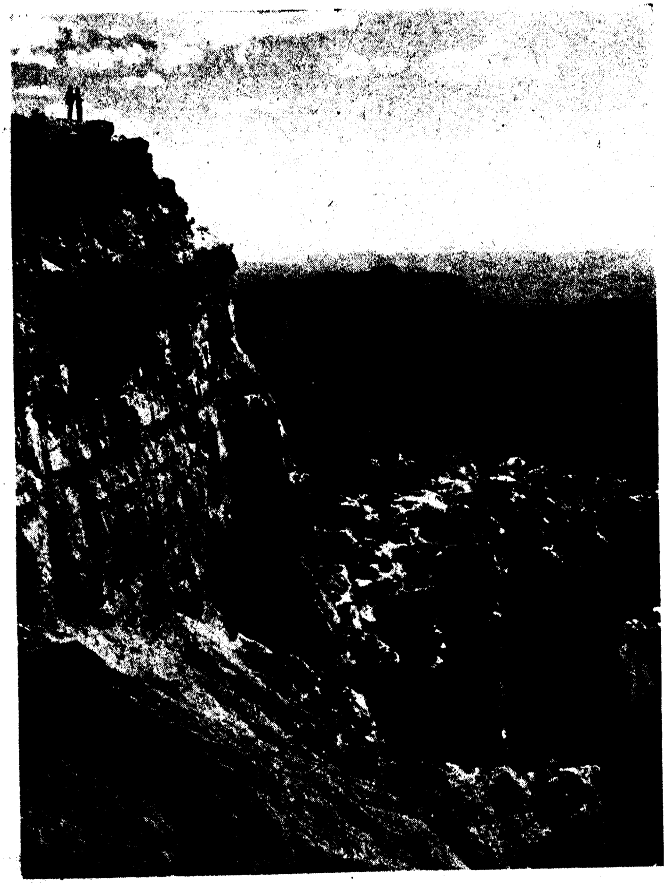


FIG. 1.—Limestones in the Wasatch formation (Eocene); typical exposure along the rim of the Markagunt Plateau. The overlying calcareous silts of the Brian Head formation have been eroded from its surface. Photographed by the U.S. Forest Service.

FIG. 2.—Brian Head formation; thin-bedded calcareous phase (the white Wasatch of early reports). Massive, pink limestone of the Wasatch formation lies about 150 feet below. Branch of Red Canyon, Garfield County, Utah.

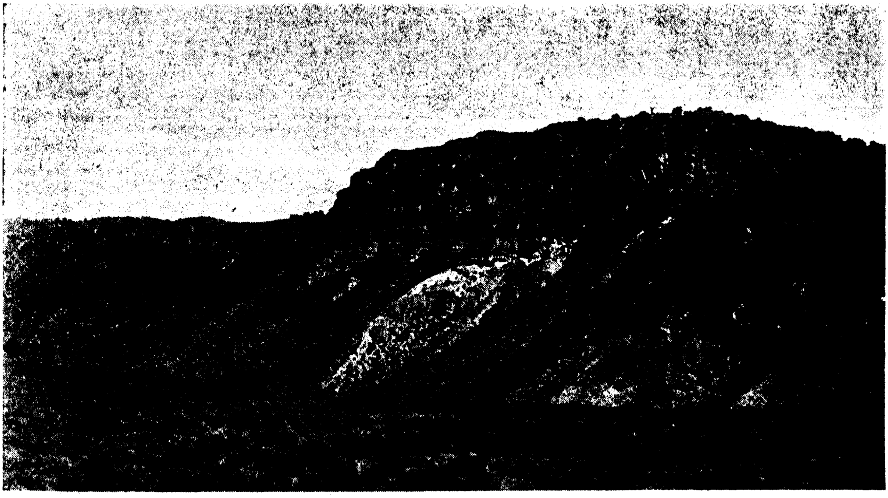


FIG. 3.—Brian Head formation and overlying andesitic lavas in the upthrown block of the Paunsaugunt fault. East Fork of the Sevier River near the head of Black Canyon.

Recent regional surveys seem to justify the recognition of the limestones, bedded pyroclastics, and conglomerates that lie between the limestone of the typical Wasatch formation and the widely spread lava flows as a distinctive unit in Tertiary stratigraphy for which the term "Brian Head formation" is here applied. A representative section is exposed at the type locality of Brian Head, a prominent projection on the western rim of the Markagunt Plateau near Cedar Breaks National Monument.

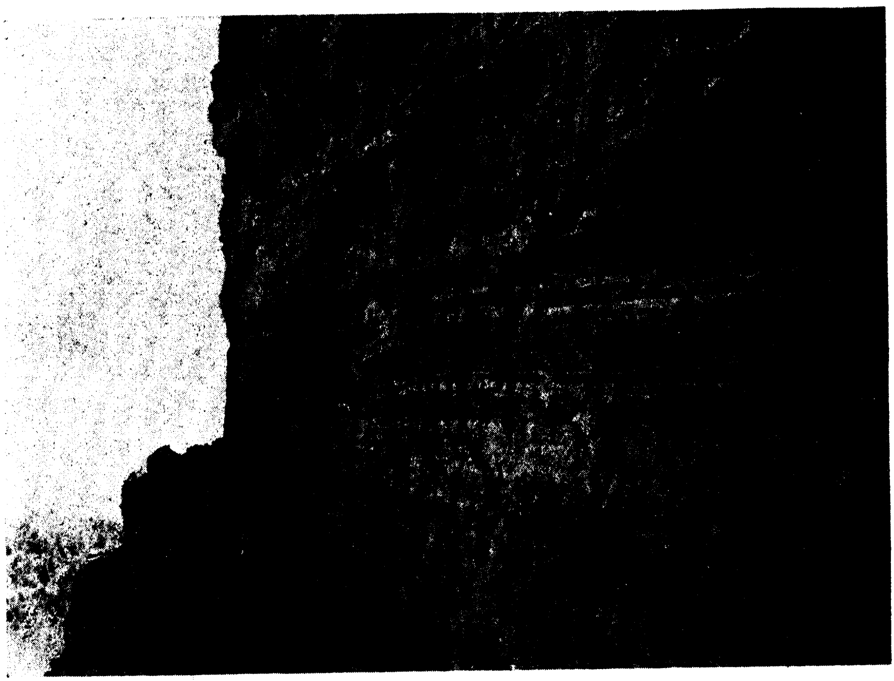


FIG. 4.—Brian Head formation: bedded pyroclastics. Face of Casto Bluff, southwest edge of Sevier Plateau.

Where fully represented the Brian Head formation includes two strongly contrasted subdivisions: a lower unit of evenly stratified, fine-grained materials, and an upper unit of coarse agglomerates. Both subdivisions vary widely in thickness, and, because of erosion or nondeposition, one or the other is absent in places. On the southern Aquarius, Paunsaugunt, and Markagunt plateaus only the lower unit is represented, and generally north of Kingston Gorge, Circleville Canyon, and Bear Valley only the higher unit.

The stratigraphic limits of the Brian Head formation are fairly well defined. Generally the contact with the underlying limestone of the Wasatch formation is marked by an abrupt change in the color, composition, and texture of the sediments and in places by an erosional unconformity. Along the west base of the Aquarius Plateau above Black Canyon a thin bed of the lower unit of the Brian Head formation rests

on a maturely eroded surface of Wasatch. In Antimony Canyon all the Wasatch and most of the lower unit of the Brian Head has been worn away, and Tertiary breccias and lavas rest on Cretaceous shales. At the upper boundary of the Brian Head formation igneous conglomerates lie unconformably beneath the lavas that in places cap the plateaus.

The sources of the materials in the Brian Head formation have not been identified. Some of the loosely compacted, regularly stratified, and dominantly calcareous beds may represent the decomposition and redeposition of the underlying compact pink limestone, but their component chalcedony and tuffaceous conglomerate are out of place in the normal Wasatch formation. The pyroclastic beds are obviously not local in origin. They are water-laid sediments derived from the disintegration of lavas that are exposed nowhere in southern Utah. The interpretation of the physical features of the Brian Head seems to in-

FIG. 5.—Brian Head formation, showing style of erosion in thin-bedded volcanic ash, tuff, and siliceous limestone. Limekiln Gulch—a tributary to the South Fork of Sevier River.

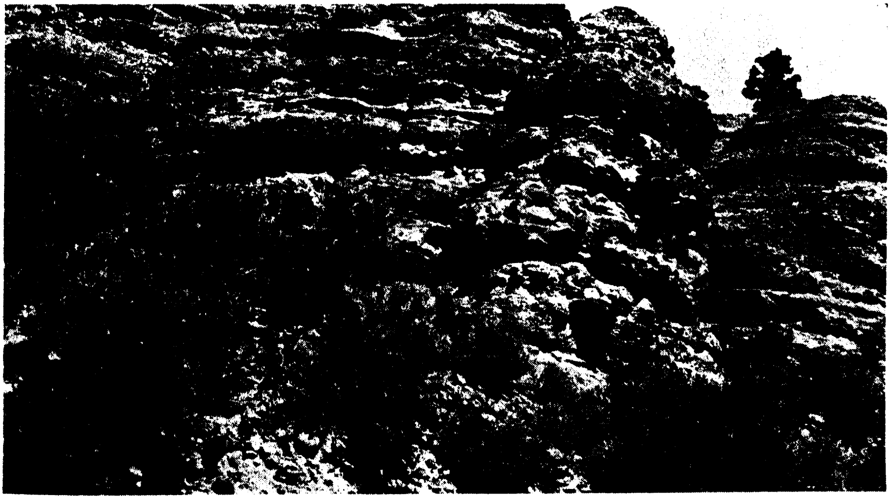


FIG. 6.—Brian Head formation: conglomerate phase. East Fork of Sevier River near the mouth of Black Canyon.

volve the assumption that lavas and calcareous sedimentary rocks in regions north or northwest of the southern High plateaus were completely broken down and their disintegration products spread over the limestone surface of the Wasatch formation.

obviously older than the Sevier River formation, which contains fossils of upper Pliocene (?) age. The formation predates the movements along the Hurricane, Sevier, and Paunsaugunt faults in late Tertiary and Quaternary times.

The outstanding features of the Brian Head formation are illustrated in Figures 1–6. The characteristics of the pyroclastic phase are shown in the representative sections (Tables 1 and 2).

TABLE 1

GENERALIZED SECTION OF THE BRIAN HEAD FORMATION AT BRIAN HEAD

	F.
Glacial deposits	
Rhyolitic lavas in thin sheets; forms top of Brian Head	
Unconformity; surface of erosion	
Brian Head formation	
5. Volcanic ash; red, highly compacted, includes tuff and ash, thin sheets of acidic lavas	6– 18
Unconformity	
4. Breccia, angular blocks of acidic lava, variable in texture, color tone, and composition; commonly 6 to 12 inches, some as much as 6 feet, in diameter. Near the top includes broken sheets of black scoriaceous andesite and lenticular aggregates of ash and lapilli. Base concealed but at one place below the breccia dense acidic rock with strong parallel horizontal joints suggests an intrusion	8–
3. Lava, acidic porphyritic, glassy; thin regular sheets marked by flow structures; breaks into sharp angular steps	0– 6
2. Volcanic ash in irregular beds, 2 to 8 inches thick; finer grained and more thinly and evenly bedded toward the top; includes lenses of coarse sandstone, siliceous limestone, clay, and chalcedony and is marked by black bands, chiefly magnetite; much of the rock is porous and in places resembles travertine and geyserite	265
1. Volcanic ash; with subordinate calcareous sandstone; gray, yellow, white, unevenly bedded; abundantly planted with pink, brown, white, and black chalcedony, distributed as nodules and lenses; near the middle continuous beds of chalcedony are 2 to 5 feet thick and as much as 500 feet long	210
Total	489–539
Unconformity?	
Limestone typical of the Wasatch formation	

In age the Brian Head formation is tentatively considered to be Miocene, but the evidence for this assignment is far from satisfactory. The rare fossil snails and fragmental turtle bones belong to genera that range throughout the Tertiary; few have been specifically determined. On the Markagunt Plateau the formation lies above rocks of Eocene—possibly middle or even late Eocene—age, and below Tertiary volcanics, which in turn underlie Quaternary basalts and glacial deposits. It is

PARUNUWEAP FORMATION

In southwestern Utah most valleys are floored with alluvial sands that record several periods of aggradation and degredation during Recent time and perhaps late Pleistocene. At heights of 10–200+feet above this friable valley fill consolidated or partly consolidated

masses of boulders, gravel, and stratified sand lie at the top or midway up on the walls of canyons, completely fill some short shallow valleys, and extend onto some low interstream divides. Within the drainage basins of the Virgin, Kanab, and Paria rivers the remnants of this material, displayed as terraces, benches, and aggregates of various shapes, occupy comparable topographic positions and are believed to represent stream-borne debris deposited during a single epoch of regional aggradation. They are classed as parts of the Parunuweap formation—a name derived from the Parunuweap valley, where typical exposures are numerous (see Fig. 7).

TABLE 2

SECTION OF THE EDGE OF SEVIER PLATEAU 2± MILES WEST OF CASTO BLUFF

	Feet
Andesite, basalt, and igneous breccia undifferentiated; forms prominent headland on rim of Sevier Plateau	600+
Brian Head formation	
8. Conglomerate of igneous fragments	90+
7. Tuff, ash, and shale, green white, gray white, fairly regular beds and long thin lenses, unsystematically hard and very soft. Includes lenses of fine quartz-quartzite conglomerate and much chalcedony. Weathers as intricately dissected badland slope or, where protected by Nos. 8 or 9, vertically grooved cliffs. Contact with Nos. 6 and 8 not definite	740
6. Volcanic ash and siliceous limestone; general green tone; hard and very soft discontinuous layers; includes rare lenses of fine conglomerate; weathers in badland forms with hardened surfaces	50–200
Total Brian Head formation	1,030+
Wasatch formation	
5. Limestone, red, sandy	15
In combination Nos. 2–5 form a series of color-banded benches.	
4. Limestone, irregular beds, gray, very sandy	40+
3. Limestone, red, poorly consolidated	34
2. Limestone, yellow, banded red and white, very friable, sandy	36
1. Limestone, pink, massive and in thick beds; includes lenses of conglomerate, made chiefly of quartzite and quartz pebbles; in nearby Casto Canyon forms vertical walls, projecting buttresses, picturesque columns, towers, statuettes, and windows. Base concealed	360
Total Wasatch formation	489+
Total thickness measured	2,119+

COMPOSITION

The Parunuweap formation includes two chief classes of sediments—conglomerates and alluvial or lacustrine silts—both of which vary widely in composition from place to place. In the Virgin and Parunuweap valleys the conglomeratic phase of the Parunuweap consists chiefly of angular slabs of gray sandstone 1–4 feet in length, partly worn pebbles of limestone, elongated iron concretions 1–12 inches in diameter, and rounded pebbles of quartz, quartzite, and chert, generally less than 3 inches in diameter. In Kanab Valley the conglomerate includes angular chunks of basalt and many mud balls. In Coal Creek Valley

FIG. 7.—Parunuweap formation (*Tp*), overlain by Tertiary (?) basalt (*B*), and underlain by the Cr taceous Waheap and Straight Cliffs sandstones, undifferentiated (*Kws*). Parunuweap Valley, 2 miles above Glendale-type locality.

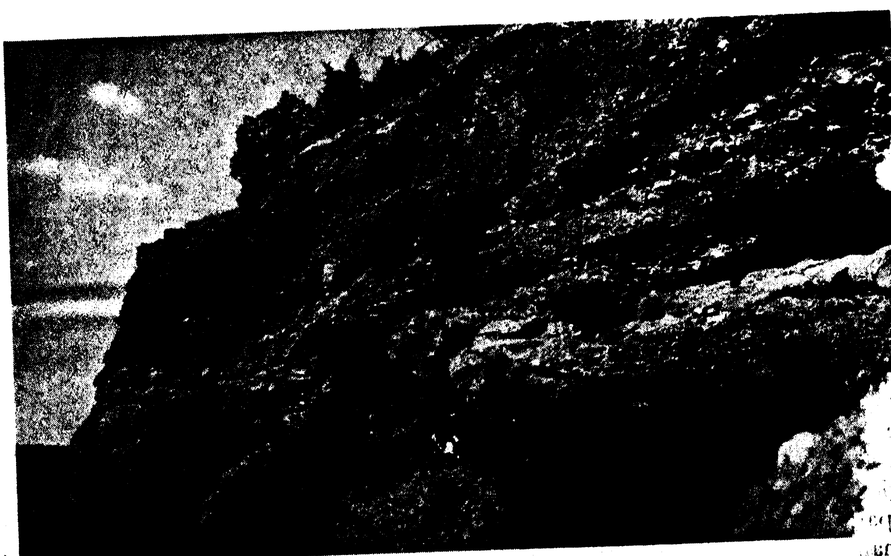
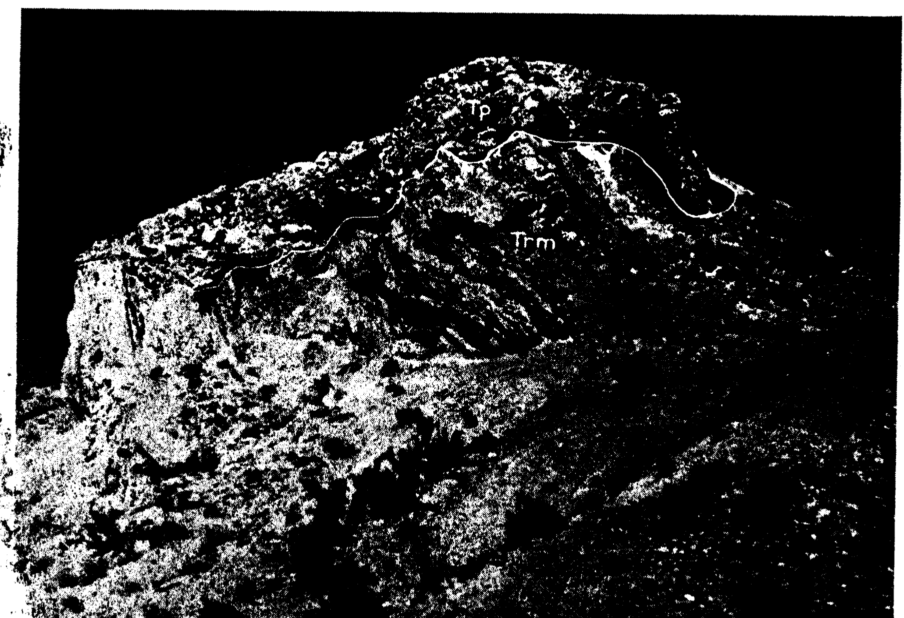


FIG. 8.—Parunuweap formation: roughly bedded conglomerate and sandstone, representative exposures in Kane, Washington, and Iron counties, Utah. Junction of federal highways 15 a Mount Carmel.

the chief components are slabs of pink and red limestone, gray sandstone, and rhyolite 1–3 feet in diameter; angular and rounded pebbles of quartzite; and chips of chalcedony. Everywhere the cement of the conglomerate is calcareous and includes enough iron oxides to produce in places a general tint of brown or yellow. Along branches of Paria River the coarse gravels have been consolidated into a caliche mass that forms persistent caps of mesas 50–70 feet above the stream bed. The lacustrine type of Parunuweap deposits, well represented along Lawrence Creek (a tributary to Coal Creek), consists of thin layers of fine-grained quartz sand and closely foliated sheets of calcareous and gypsiferous silts. The slight unconformities between groups of beds indicate seasonal pauses in deposition (see Fig. 9).

FIG. 9.—Parunuweap formation: lacustrine and fluvial silts. Lawrence Canyon, a branch of Coal Canyon.

The Parunuweap formation rests on a surface of erosion developed on rocks of



…. 10.—Parunuweap formation (*Tp*), resting on truncated beds of the Moenkopi formation (*Trm*), …lted in the Kanarra fold. Mouth of Coal Canyon, near Cedar City.

various ages; from place to place it overlies indifferently the Triassic, Jurassic, and Cretaceous formations. Its upper contact is equally well marked; beds in the Parunuweap are unconformably

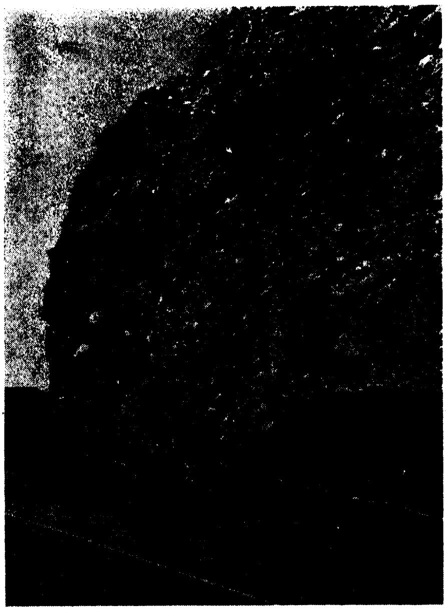


FIG. 11.—Parunuweap formation: conglomerate resting unconformably on Triassic shales. Virgin River Valley near Grafton.

overlain by sheets of basalt or by talus, stratified sand, and gravel of very recent age. As defined by these limits, the existing exposures are 30–80 feet thick. Before erosion had removed its upper part, its thickness in places probably exceeded 100 feet (see Figs. 10, 11, 12).

ORIGIN AND AGE

The numerous remnants of the Parunuweap formation in southwestern Utah are restricted to valleys, low divides, and shallow basins on the plateau tops. They consist of materials that characterize shattered ledges, alluvial fans, talus slopes, and the disintegrated rock on interstream spaces. From nearby areas they seem to have been transported by strong streams of fluctuating speed and volume and dumped into valleys or other available depressions.

Though the Parunu eap formation has yielded no fossils of diagnostic value its stratigraphic position among other formations of late Tertiary and Quaternary times makes a tentative assignment to the Pliocene seem reasonable. It is older than the Quaternary basalts, the fossiliferous Pleistocene sediments, and

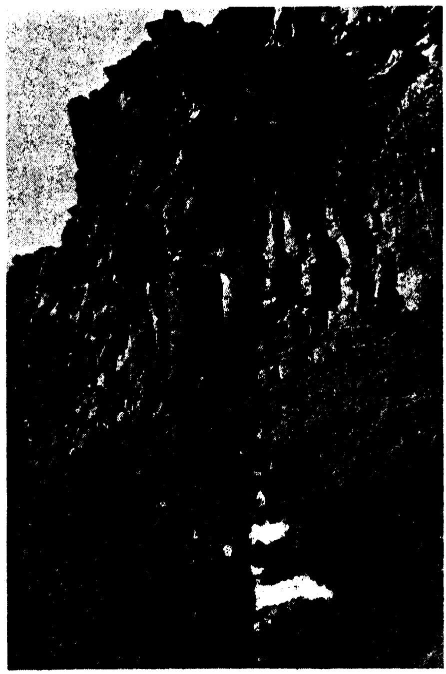


FIG. 12.—Parunuweap formation and overlying Tertiary (?) basalt; the partly eroded gravel fill of an ancient valley overrun by lava flows. Wall of La Verkin Canyon 2 miles south of Toquerville.

the regional uplifts and faults that out- the High Plateaus. It is younger than the widespread rhyolites, andesites, nd latites of probable Miocene age and pre- ates part of the deep stream trenching

that characterizes the Canyon Cycle of erosion. In distribution, composition, and general physical aspects it closely resembles the materials of late Pliocene or early Pleistocene age that in central Utah constitute the Sevier River formation.[4] Also in stratigraphic position and regional relations they are comparable to the Pliocene deposits that in the Navajo country lie on a surface of erosion developed on Cretaceous, Jurassic, and Triassic rocks.[5]

THE SEVIER RIVER FORMATION

In southwestern Utah rocks assigned to the Sevier River formation of late Pliocene or early Pleistocene age form part of the faulted fanglomerates that extend westward from the base of the Sunset Cliffs and are well exposed along the South Fork of the Sevier River between Hatch and Panguitch, where they underlie Quaternary basalts. The deposits make up partly consolidated, roughly stratified beds of conglomerate, sandstone, and arenaceous shale, of which the most conspicuous components

FIG. 13.—Sevier River formation: view typical of exposures along the South Fork of the Sevier River.

are subangular fragments of andesite and other igneous rocks 3–10 inches in diameter. The finer-grained materials are quartzose gravel and sand, which include some volcanic ash and clay shale (see Fig. 13). Like the Parunuweap formation, with which it is tentatively correlated, the Sevier River formation is not continuous over large areas. It occupies streamways and other depressions into which materials from surrounding regions have been carried.

[4] Eugene Callaghan, "Preliminary Report on the Alunite Deposits of the Marysville Region, Utah," *U.S. Geol. Surv. Bull. 886* (1933), pp. 11–101.

[5] Gregory, "The Geology of the Navajo Country," *U.S. Surv. Prof. Paper 93* (1917), p. 121; Howell Williams, "Pleistocene Volcanoes of the Navajo-Hopi Country," *Bull. Geol. Soc. Amer.*, Vol. XLVII (1936), pp. 111–72; J. T. Hack, "Sedimentation and Volcanism in the Hopi Buttes, Arizona," *Bull. Geol. Soc. Amer.*, Vol. LIII, No. 2 (1942), pp. 335–72.

LANDSLIDE IN ZION CANYON, ZION NATIONAL PARK, UTAH

RUSSELL K. GRATER
Rocky Mountain National Park, Estes Park, Colorado

ABSTRACT

The existence of a large landslide that blocked Zion Canyon and created an extensive lake has been demonstrated through mapping of the slide and discovery of the old lake deposits. The slide came from Sentinel Mountain on the west side of the canyon. It was caused by an extensive fracture and joint system, water-saturated shale beds underlying the massive Navajo sandstone, and a regional dip of 1°–3° to the northeast. Erosion of the slide debris is still very rapid, as is shown by the major slides of 1923 and 1941 that temporarily dammed the Virgin River.

INTRODUCTION

For some time it has been recognized by geologists that the level floor of Zion Canyon was probably caused by a major landslide several thousand years ago that dammed up the Virgin River and caused sedimentation in the still waters behind the barrier. In recent years rock slides of sufficient size to destroy whole sections of the valley road and to dam the river have brought a renewal of interest in this phase of the canyon story. In an attempt to work out the details of what took place, the writer has spent as much time as possible during the last two years investigating and interpreting the available evidence.

GEOLOGICAL REVIEW OF THE AREA

The Markagunt Plateau is the extreme southwestern portion of the Plateau Province in southern Utah. To the west it is bounded by the Hurricane fault, with a maximum throw of approximately 6,000 feet. Its eastern edge is marked by the Sevier fault, with a displacement of approximately 2,000 feet. Starting from near sea-level, the plateau has been raised until its highest point is now over 11,000 feet above the sea. During the uplift the southern end of the block rose more than the northern portion, creating a regional dip to the northward. In similar fashion the western edge of the block exceeded the uplift along the Sevier fault, causing the rock strata to dip to the eastward from 1° to 3°. The greater part of the northern end is covered by lavas.

The uplift of the Markagunt Plateau was a slow and intermittent movement. As the highlands developed, conditions for erosion became more favorable, and the established streams became powerful scouring agents as their gradients steepened. Thus were carved the many sheer-walled canyons so characteristic of the Zion region. During this period of uplift a striking regional system of joints was developing. The pattern of this system is very uniform, the major fractures trending north-northwest throughout the plateau. These vertical cracks have a strong influence upon the drainage system of the region, as the streams tend to follow along these zones of weakness.

STRATIGRAPHY

Of the rock column exposed in the Markagunt Plateau, only those strata of Triassic and Jurassic age are of importance in this paper. The Triassic is made up of the sandstone, shale, gypsum, and limestone of the Moenkopi formation (approximately 1,800 feet, Lower Triassic), the conglomerate of the Shinarump

formation (approximately 85 feet, Upper Triassic), and the sandstone, shale, gypsum and limestone conglomerates of the Chinle formation (approximately 1,000 feet, Upper Triassic). With the exception of the Shinarump and one member of the Chinle, these rocks are rather weak and are readily eroded. The Jurassic is composed of the Wingate (approximately 150 feet, Lower Jurassic?), a strong, cross-bedded sandstone; the Kayenta (approximately 40 feet, Lower Jurassic?), a series of coarse- and fine-grained sandstones intermingled with shale beds; the Navajo (approximately 2,100 feet, Middle Jurassic?), a massive, fine-grained, friable, quartz sandstone; and the Carmel (approximately 250 feet, Upper Jurassic), a group of hard limestone beds separated by thin beds of limy sand. Of these formations, the Wingate, Navajo, and Carmel are of sufficient strength to resist rapid erosion, while the Kayenta is weak.

STATEMENT OF THE PROBLEM

To the visitor viewing Zion National Park for the first time, there is a tantalizing similarity in shape between the sheer-walled gorge of Zion Canyon and the Yosemite Valley in Yosemite National Park, though the colors of the two areas are decidedly different. The impression is so pronounced that the explanation of this similarity is a daily task for the members of the naturalist staff of the park. To do this adequately has been a bit difficult. It was easy to show that glaciation did not occur in Zion Canyon and that the apparent "ice lines" on the cliffs were really only dune deposits of the Navajo formation in cross-section. Even the sheer walls and U-shaped valley could be explained by pointing out how the fracture system, plus the action of natural arches created by seeps at the base of the cliffs, worked together to keep the walls relatively vertical. But the most difficult problem was to explain the pronounced flatness of the canyon floor (Fig. 1). By comparison, the other major canyons of the region have floors that are only roughly flat. It was to develop a better understanding of this problem that the study was inaugurated.

GENERAL FIELD EVIDENCE

A preliminary survey of the valley disclosed that the flat floor is primarily of water-laid clays and fine sands. Zion Canyon has been carved through the erosive power of the Virgin River, whose spring floods allow very little deposition of the flood-plain type. Inasmuch as the vast deposits in the valley floor could have formed only in relatively still water, the possibility of a flood-plain accumulation could be quickly discarded. Since still waters were required for the deposition of the valley materials, it is clear that a barrier must have been created across the valley at some time which impounded its waters.

The belief that a major barrier once existed near the mouth of the canyon was soon confirmed. There was no difficulty in locating a slide mass which blocked the stream or in determining its minimum extent. Reconnaissance showed that the slide had come from the east face of Sentinel Mountain (see Fig. 2), where some highly shattered rocks remain to show the source of the material. The slide itself is approximately $1\frac{1}{2}$ miles in length and $\frac{3}{4}$ mile in width at its widest point. Near its crest is a vast field of huge boulders of the white Upper Navajo sandstone, now almost a half-mile from the base of the nearest cliff. Behind the crest of the slide is a small valley with gently sloping terrain to both south and north where small intermittent streams

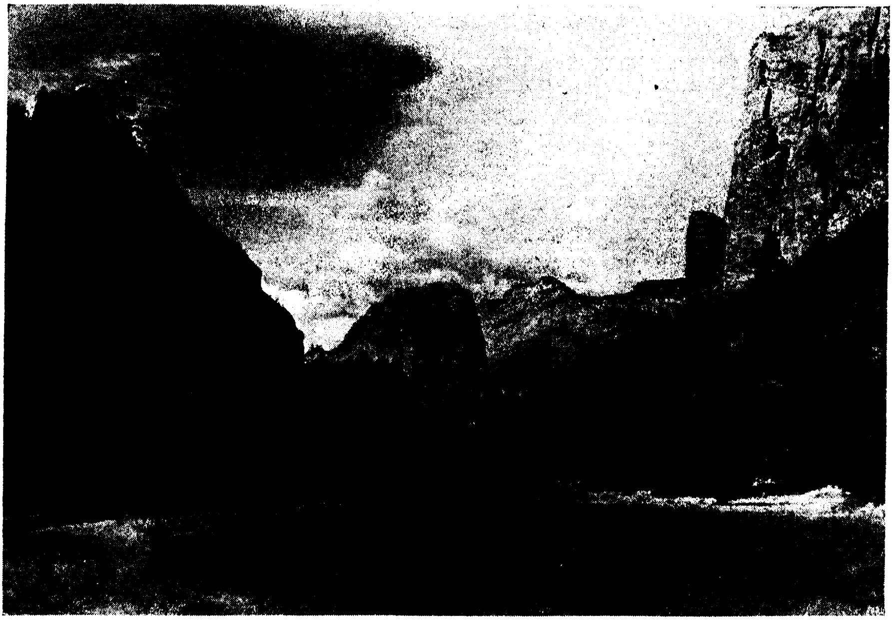


FIG. 1.—Zion Canyon near Zion Lodge. The characteristic flatness of the valley floor is well illustrated in this sector.

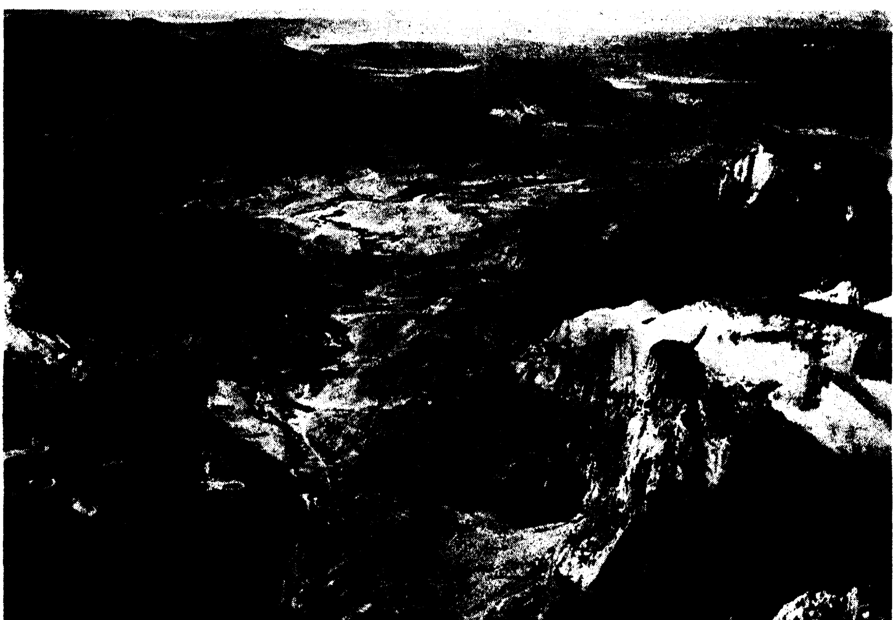


FIG. 2.—Aerial view showing the slide area from point above the Court of the Patriarchs. Note the highly shattered area around the east face of the Sentinel.

have removed part of the loose debris. In the southern portion of the slide area, the Virgin River has removed all but one remnant from near the stream course, this one remnant apparently being from a small landslide similar to those that blocked the river and destroyed portions of the canyon highway in 1923 and 1941 (see Fig. 3). In both of these instances some of the loose debris of the older slide mass moved down, damming the Virgin River for a short period of time.

The cause of the major slide is immediately apparent from a study of the topography and rock strata. As has been pointed out, there is pronounced rock fracturing in places, with cracks extending from the uppermost formations downward through the exposed beds and into the underlying rocks. One such zone of fracture extends from the Wildcat Canyon–Great West Canyon area through the Court of the Patriarchs and into or past the Sentinel (see Fig. 4). Because of this fracture system large blocks became capable of movement away from the canyon wall. Couple with this the fact that the weak shale beds underlying the massive Navajo sandstone are dipping to the eastward, and the cause of the slide becomes clear. The drainage developed in this area indicates that these fractures have been a strong factor in developing the present topography.

Following the mapping of the slide, an examination was made of the valley between the Court of the Patriarchs and the Temple of Sinawava—a sector about 5 miles in length. Remnants of old lake fill in the form of well-bedded, finely laminated clay deposits were uncovered in Birch Creek Canyon (see Fig. 5), at various places along the sides of the valley approaching Zion Lodge, in a small unnamed alcove opposite Zion Lodge, at the Grotto Campground, and near the foot of Angel's Landing. The small alcove across from Zion Lodge afforded the best exposure of the bedding, and a survey was made to determine the elevation of the uppermost bedding plane. This was found to fall on the 4,320-foot contour, while the elevation on the valley floor was 4,269 feet. The river level at this

FIG. 3.—View of the 1941 slide that destroyed a portion of the valley road system. This debris came from the old slide originating on Sentinel Mountain.

point is approximately 4,258 feet. Using the 4,320-foot contour as the maximum height at which known still-water deposition had occurred (highest known lake level), a chart was made showing the approximate extent of the lake (see Fig. 6). Obviously the ancient lake created by the slide from the Sentinel must have extended up the canyon at least as far as the Weeping Rock area.

The park engineer uncovered additional important evidence of lake fill in

test pits for the site of a small suspension bridge across the Virgin River opposite Zion Lodge. The first pit was dug approximately 25 feet from the nearest talus slope on the west bank of the river. The first 4 feet were in sand, gravel, and clay. Then came 17 feet 7 inches of blue clay. At this point a fine muck was encountered, and test rods thrust down into this oozy mass failed to reach solid bottom. On the east bank of the river a second test pit was dug at a distance of 150 feet from the initial pit. The first 8 feet of the material was of sand, gravel, and some clay. Then followed 12 feet of blue clay. At this point the digging was terminated, and test rods verified that fine muck was only a short distance beneath the clay deposit. From these test pits it is clear that the level valley floor is underlain with fine clays and silty materials that were obviously deposited in still or relatively still water.

FIG. 4.—Aerial view showing southeastern portion of the Great West Canyon-Wildcat Canyon area. Fracture zone shown runs directly past the West Patriarch on either side.

RECONSTRUCTION OF PROBABLE EVENTS

From the evidence obtained, the probable events leading up to the present-day valley can be reconstructed. At one stage in its history the Virgin River had cut down to the Springdale member of the Chinle formation in the vicinity of the Court of the Patriarchs and through it to the southward. Then came a tremendous slide that blocked the entire canyon. From the formations involved it seems probable that water, fed into the fractures, filtered downward until the shale beds on top of the Springdale member were thoroughly saturated, creating

a "greased skid" for the great mass of overlying rocks. As the slide came down into the canyon it fanned out, filling in the gorge from the present location of the Zion Canyon–Mount Carmel Road junction to the west side of the Court of the Patriarchs (see Fig. 7 and area on Fig. 6). The crest of the slide was directly opposite the Sentinel and shoved a great mass of debris high up on the slope beneath the Twin Brothers and Mount Spry. That the crest of the slide was at least 500 feet above the river-level is shown by the fact that the river was then around the 4,100-foot level, and the present high points of the slide on opposite sides of the river are 4,712 feet and 4,703 feet, respectively (see Fig. 7). Erosion has undoubtedly lowered the crest, although not to any great extent. The blocking of the river formed a large lake that filled the canyon probably back as far as the Narrows at the foot of the Mountain of Mystery. In this still water, silt and clay muds carried by the old Virgin River were dropped, forming extensive beds of clay underlain with oozy muck. In the meantime, the impounded waters finally began to pour over the top of the slide barrier. The lowest point along this barrier was on the east side of the canyon below the toe of the slide. Probably the initial erosion was rapid, as the waters had a tremendous fall and had only loose debris to move. It is believed that the maximum lake elevation was maintained for only a comparatively short time and so left no mark of its position. After this, the lowering of the lake level by removal of the slide debris was probably slower, allowing extensive deposition to continue in and around such still-water areas as Birch Creek and Emerald Pool canyons as well as behind the rock dam itself (see Figs. 8 and 9). The final draining of the lake and the exposure of the clay deposits was probably a relatively slow process, as the deeper the stream ate into the slide debris, the more its gradient was reduced and the slower it cut. This allowed for partial consolidation of the underlying clays in the lake fill, and their protection against rapid erosion by a heavy vegetative

FIG. 5.—Finely laminated clay deposits in Birch Creek Canyon.

cover that came in over the valley floor. Today, at least 76 feet of this lake fill have been removed at its lower end, yet extensive deposits remain that are virtually untouched.

There has been much speculation as to how rapidly the slide debris has been removed and will be removed in the future. Some conception of this can be obtained by records kept since the park was established in 1919. Since that date there have been two major slides involving the old slide debris. Both slides, incidentally,

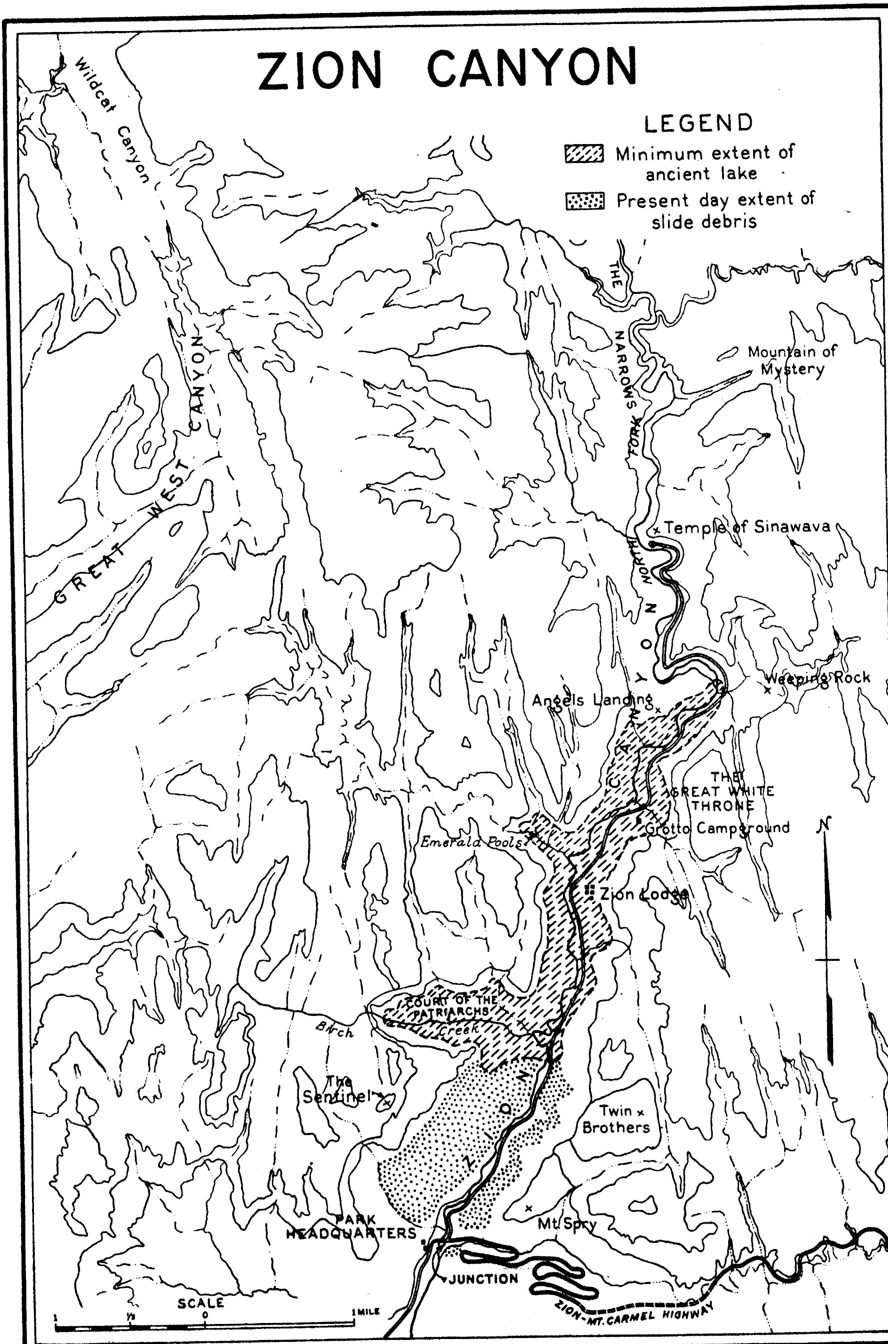


FIG .6.—Map of Zion Canyon showing the minimum extent of the ancient lake and the present-day extent of the slide debris.

FIG. 7.—View from above the junction of the Zion Canyon–Mount Carmel Highway looking up the main Zion Canyon. The extent of the slide debris and its highest elevations on both sides of the valley are clearly evident.

FIG. 8.—Clay deposits along the highway near the Court of the Patriarchs

occurred because the saturation by water of the shale beds overlying the Springdale member of the Chinle allowed the entire mass to move. No figures are available as to the amount of material that came down and blocked off the river in 1923, but the 1941 slide was estimated to have comprised at least 150,000 cubic yards of rock debris. Thus, in twenty years, the river has removed a great quantity of the old slide. Using this as a gauge, the rate of removal and erosion has been rapid, and at most only a few thousand years can have elapsed since the slide occurred.

FIG. 9.—Clay deposits laid against the slide debris as shown in the road cut at the Court of the Patriarchs. Dotted line shows outline of slide.

ACKNOWLEDGMENTS.—In these studies the writer was aided by constructive suggestions by Dr. Herbert E. Gregory, of the United States Geological Survey; Dr. John C. Hazzard, of the Union Oil Company; and Messrs. Walter Buss and Otto Cross. Assistance in obtaining the elevations in the valley was given by Chief Ranger Fred Fagergren.

VELOCITY AND LOAD OF A STREAM

TERENCE T. QUIRKE
University of Illinois, Urbana

ABSTRACT

Certain modern geology textbooks confuse the power with the force exerted by streams. It is desirable to distinguish clearly between the "capacity" of a stream to transport a load and its "competence" to shift masses of specific size and between Gilbert's definition of "capacity" of a stream and its potential energy. Some writers repeat an ancient mistake in relating capacity of a stream to its velocity. New errors follow the application to stream phenomena of empirical formulas based upon experiments in flumes, contrary to Gilbert's warning.

INTRODUCTION

Many textbooks carry the following statement: "The transportation power of a stream varies as the sixth power of the velocity." In several cases[1] this statement is followed by such an explanation that the expression "transportation power" may be interpreted properly to mean the *force* just great enough to shift a mass of some specific size or weight.[2] Other textbooks[3] are so ambiguous, if not incorrect, that one may well suppose that the "power" of transportation is to be understood literally in its technical meaning. This is especially applicable to those books written for students of engineering or other applied sciences. The writers of still other texts[4] are completely misled by the old statement and state that, if the velocity of a stream is doubled, it can carry sixty-four times as much as it carried before. The fact that professional geologists have been misled by the statement proves that it should never appear in textbooks. This is the same error that Gilbert[5] objected to in 1914, although it does not appear to have been current in textbooks of that time. Nevertheless, as far as this writer can find, practically every writer who referred to the "power" of transportation, found it necessary to explain what he meant by the term if he were to be clearly correct in meaning.[6]

Several writers,[7] presumably recognizing these confusions, have refrained

[1] T. C. Chamberlin and R. D. Salisbury, *Geology* (1904), Vol. I, p. 109; W. H. Hobbs, *Earth Features and Their Meaning* (1912), p. 159; L. V. Pirsson and Charles Schuchert, *Textbook of Geology*, Part I (2d ed.; 1919), p. 42; W. J. Miller, *Elements of Geology* (2d ed.; 1939), p. 109.

[2] Grove Karl Gilbert, "The Transportation of Debris by Running Water," *U.S. Geol. Surv. Prof. Paper 86* (1914), p. 16. Gilbert states that Hopkins formulated the law in these words: "The moving force of a current, estimated by the volume or weight of the masses of any proposed form which it is capable of moving, varies as the sixth power of the velocity."

[3] R. S. Tarr and Lawrence Martin, *College Physiography* (1918), p. 113; H. Ries and T. L. Watson, *Elements of Engineering Geology* (2d ed.; 1930), p. 131.

[4] A. W. Grabau, *Textbook of Geology* (1920), p. 46; Pirsson and Schuchert (rev. by others), *Textbooks of Geology*, Part I (3d ed.; 1929), p. 50.

[5] P. 16 of ftn. 2 (1914).

[6] Not all these explanations are beyond criticism; one author discusses velocity and transportation in such words as to make equivalents of force, power, and velocity.

[7] R. T. Chamberlin and Paul MacClintock, *Chamberlin and Salisbury's College Geology*, Part I (2d ed.; 1933); Chamberlin and Salisbury, pp. 103–4 of ftn. 1 (1904); W. H. Emmons, G. A. Thiel, C. R. Stauffer, and I. S. Allison, *Geology* (2d ed.; 1939), p. 97. These authors cite relation between bottom velocities and the size of particles moved—i.e., competence.

from any quantitative ratio between velocity of the stream and the load it does or may carry. W. B. Scott[8] says: "The mechanical work of a river is . . . . dependent upon the velocity of the current, varying directly as the square of that velocity."

This statement may be applied to any part of a river's work which depends upon the locally effective velocity, because kinetic energy is equal to one-half the mass multiplied by the velocity squared —kinetic energy being the ability to perform work. Inasmuch as a stream's kinetic energy varies directly as the square of its effective velocity, its ability to move a load appears to have the same relation. This shifts the emphasis from the size of the largest particle moved (competence) to the total load carried, and it makes clear the basic reasoning. Similarly, Pirsson[9] stated: "The corrasive power of a current varies as the square of the velocity, with equal size and distribution of particles," thus relating power and energy (in this case implied as the ability to do *work* of corrasion) with velocity.

Unfortunately, Gilbert[10] defined "capacity" in such terms that it is not proportional to either discharge, slope, or potential energy of the stream and is not statable in terms of energy. Capacity, as defined, does not measure economy of energy but merely relative accomplishment in respect to any condition selected for comparison. Consequently, "capacity," being a product of many other factors besides the kinetic energy of the stream, does not necessarily vary according to the square of the velocity. On the other hand, the potential work of the stream, including potential load transportation, must vary according to the square of the theoretical velocity.

CAPACITY AND COMPETENCE

There is a common confusion in terminology and meaning between the "capacity" and the "competence" of a stream. Gilbert[11] clearly distinguished between the two: The maximum load a stream can carry is its capacity; the size[12] of the particles of debris a stream can move is a measure of its competence. Even very careful and authoritative writers fall into error, unless this critic himself is much mistaken. P. G. Worcester,[13] referring to the statement of Gilbert, states that

> the capacity of a stream to move its bed load varies with the 3.2 power to the 4.0 power of the velocity. Using the latter figure, it is equivalent to saying that if the velocity is doubled the stream's power to move objects on the bottom is increased 16 times. This principle applies primarily to the size of the particles moved. . . .

This clearly is a case of confusion, because where Gilbert says "capacity" he does not mean "competence," and the statement is applied by Worcester not to "capacity" but to "competence." O. D. von Engeln[14] says:

> The transporting power of a stream for such tools is proportional to the sixth power of the

[8] *An Introduction to Geology* (2d ed.; 1927), p. 136.

[9] *Introductory Geology* (2d ed.; 1924), p. 40. Unfortunately, Pirsson follows faulty reasoning, and his other conclusions noted below are not correct.

[10] P. 36 of ftn. 2 (1914).

[11] *Ibid.*, p. 35.

[12] The size of a particle may be measured as the cross-sectional area upon which the current may impinge. It is assumed that the particle is equidimensional and that its volume is proportional to its diameter. In making comparisons it is assumed that the specific gravity of all particles is the same and that other factors, such as roughness of the particle and of the bed of the stream, are similarly constant. This note applies to every case where the term "size of particle" occurs in this paper.

[13] *Textbook of Geomorphology* (1939), pp. 164–65.

[14] *Geomorphology* (1940), pp. 175–76.

current. . . . . The derived power of corrasion is computed to vary as the square of the velocity. The indicated tremendous increase in corrasive power resulting from slight increases in velocity of flow is for geomorphology a relation of first importance.

Apparently, the foregoing statements are correct in intention, although literally ambiguous. For "transporting power of a stream for such tools" we may write "transporting *competence* of a stream for such tools." This writer presumes that "corrasive power" in this connection means the power of a stream to corrade its bed by shifting its load downstream. Power is the rate at which work is done; in this case, work may be expressed as the movement of points of applied force against resistance. Evidently the work involved is overcoming friction, only a fraction of which is against the bed of the stream. Nevertheless, that fraction is presumed to vary according to the square of the effective velocity of the stream, simply because it is a form of work. It may be noted that effective velocity refers to the velocity of the stream responsible for the energy applicable and that in these arguments the potential energy of each unit of a stream's load is neglected, although in all cases it is real and, on steep stream beds, where boulders roll along, it must be appreciable. However, the present writer infers that von Engeln had in mind transporting competence (proportional to V^6) rather than "corrasive power" (proportional to V^2?) when he wrote the last sentence quoted; if so, he fell into the same old confusion between competence and capacity. Actually, neither competence nor capacity, by definition, involves power, the rate at which work may be done; of the two, only capacity involves energy, the ability to do work. However, the geomorphologist is usually interested in power, the rate at which geological work—that of corrasion, for instance—is done. Of course, capacity may be qualified by a time factor, but competence applies only to the maximum size of the particle moved and not at all to the speed at which it is shifted.

Other related errors are attributed to R. F. Flint,[15] who writes:

Because friction between water and channel slows a stream down, velocity is greatest in channels with the smallest area in proportion to the volume of water. Deep narrow channels therefore give greater stream velocity than broad shallow ones.

Both sentences need correction. Obviously, streams flow fastest in the most constricted channels, which are the "bottlenecks"; but that is not the result

[15] R. F. Flint, C. R. Longwell, and Adolph Knopf, *Outlines of Physical Geology* (1934), p. 37.

The second edition of the foregoing work, dated 1941, has a greatly condensed treatment of the same topic. However, it is still subject to criticism. P. 71 states: "Under ideal conditions, doubling the velocity may (1) increase abrasive power about 4 times; (2) increase the capacity to transport rock fragments of a given size by as much as 32 times; or (3) increase the volume of the largest piece of rock the stream can push along its bed by as much as 64 times." Items (1) and (3) may pass, but item (2) has no significance. Suppose we consider the size of fragment for which the stream is competent at any velocity. The load of that size moved is very small. Now let the velocity be doubled. The quantity of fragments of the same size now transported is very great, certainly 32 times as great as before and probably much more. Suppose we choose a piece larger than could be moved at the first velocity. The amount carried first is zero; as soon as the velocity becomes competent to move any pieces of that size, the increase in transportation becomes infinite. If, on the other hand, we choose small material for which the stream in the first case is fully competent, then doubling the velocity of the stream will not increase the load of that small size very greatly. Of course, it is possible also to choose a size beyond the competence of the stream at either velocity. In general, there must always be a size for which the increase is as V^5, as stated above; but for all other sizes the increase varies as between V^∞ and zero. Consequently, the statement, item (2) above, is misleading and ought to be clarified or left out.

of low friction. It is not necessarily true that deep, narrow channels promote greater stream velocity; it depends upon how deep and how narrow they are in proportion to the cross-sectional area of the stream. The point depends upon the "hydraulic radius," the ratio of cross-sectional area to length of wetted perimeter. Under "laws of erosive power," transporting "power" is made equivalent to competence; and that is evaluated in the following words: "The maximum diameter of the individual rock fragments a stream can move varies as the square of the velocity." This is equivalent to saying that the competence varies as the sixth power of the effective velocity, which is accepted as correct. However, the author continues: "The abrasive power is said to vary between the square and the sixth power of the velocity." This appears to be a repetition of a statement by Pirsson[16] based upon a simple error in argument. He assumes that, inasmuch as the maximum size of a particle which can be moved varies as the sixth power of the velocity, the momentum attained by such a particle also varies in the same ratio. This is not the case. Momentum is mass multiplied by velocity. The velocity of the maximum moving mass barely exceeds zero; and, of course, the product of such a velocity with any mass is small. Never, under otherwise constant conditions, do the largest particles move on the bottom as fast as the stream which moves them.

The meaning implied in the expression "abrasive power" is not clear, but it is interpreted as an expression for the ability of a stream to corrade its bed. It is assumed that each moving particle scraping the bed of a stream corrades in proportion to the product of its mass and its velocity. It is also assumed that the bed is not protected from abrasion by a sheet of sediment and that the load available is the maximum load which can be carried by the stream. The size of the largest particle which can be moved, of course, is a measure of the competence of the stream; but there is no indication as to the speed at which it may be moved. The term "abrasive power" seems to lack both definition and probable usefulness.

The relative competence of a stream from place to place is indicated by variation in size of the pieces transported in different parts of its course. A table showing such behavior of the Mur River is presented by the Minnesota group[17] as a quantitative illustration of the fact that the diameter of the particles in a stream's load becomes smaller by corrasion during transportation. Perhaps the table may be regarded better as an illustration of the sorting effect of a stream. Pieces too large for the competence of the local current must be left behind. The significance of competence in general appears to deserve more specific recognition.

W. H. Twenhofel[18] follows Gilbert in distinguishing clearly between competence and capacity but states that "the ratio of capacity to velocity increase ranges between the 3.2 and the 4th power of the increase." The present writer has not located the authority for this statement. Thereafter Twenhofel quotes Gilbert (p. 11) to the effect that capacity varies as between the 3.2 and 4.0 power of the velocity with proper modifications. Apparently, these figures have been or are about to be accepted by

[16] Pirsson, pp. 41–44 of ftn. 9 (1924). The writer thanks Dr. M. King Hubbert for the note that Pirsson in this point follows Joseph LeConte.

[17] Emmons, Thiel, Stauffer, and Allison, p. 98 of ftn. 7 (1939).

[18] *Principles of Sedimentation* (1939), p. 193.

students of sedimentation, judged by other references quoted, although Gilbert[19] himself warned that "they do not permit an estimate of a river's capacity to be based on the determined capacities of laboratory streams."

It is understood that a natural stream applies a relatively small portion of its available energy to load transportation. Computations based on the behavior of the Mississippi River at Columbus, Kentucky, following formulas by Captain A. A. Humphreys and Lieutenant H. L. Abbot,[20] show that the actual velocity of the river at that location is 90 per cent of the theoretical velocity, on the assumption of no obstructions to free flow. Apparently friction, owing to bends and changes in cross-sectional areas and shapes and because of the load, dissipated enough energy so that the velocity was reduced by 10 per cent. Only part of that computed reduction may be charged to frictional resistance of load transportation, so that only a small part —much less than 10 per cent of the kinetic energy of the stream—is applied to carrying the load. In cases of shallow, turbulent, and heavily loaded streams, of course, a larger proportion of the total available energy would be used up in shifting the load. Nevertheless, whatever work a stream does must reduce to that extent its remaining energy and, accordingly, its effective velocity. Consequently, the actual effective energy of a stream carrying a load may be expressed in terms of $V-x$, in which V equals the velocity of free flow unimpeded by any load, if all other losses of energy may be disregarded in this connection, and x represents the reduction of velocity due to the load.[21] If the load carried is very large, as in the case of a shallow stream supplied with abundance of load smaller in size than the maximum for which the stream is competent, then the factor x must be relatively large. In such a case, capacity (C) may vary as $(V - x)^{3\pm}$, whereas it varies to a degree less than V^2. In large, deep streams the factor x is relatively small, and then capacity varies almost as V^2. In small streams, in laboratory troughs, and in miners' sluices, both the frictional drag and the energy consumed by load transportation are large and become highly significant, reducing sharply the actual velocity of the loaded stream, as any placer miner knows. Consequently, the load carried varies according to a much higher power of the measured velocity. In each case the kinetic energy of the stream depends upon V^2, that is, the velocity of free flow unimpeded by any load. The more efficient the stream as a transportation agent, the greater is the factor x; also the lower the value of $V - x$ and the higher the synthetic exponent of $(V - x)$ which must be applied to relate "velocity" to capacity. Evidently, it is the factor $(V - x)$ which is recorded as the measured mean velocity in streams. Consequently, the measured velocity of a loaded stream is never great enough to be strictly pro-

[19] P. 16 of ftn. 2 (1914).

[20] "Report upon the Physics and Hydraulics of the Mississippi River," *Corps of Engineers, U.S. Army, Prof. Papers* (Washington, 1876), data on pp. 605 and 335; formula (41), p. 332. Computed velocity is 7.6 ft./sec; measured mean velocity was 6.9 ft./sec.

[21] Similarly, Humphreys and Abbot (*ibid.*, p. 337) note that in natural channels "a certain part of the actual slope is consumed in overcoming the resistances opposed by the inequalities of cross-section." They deduce further: "A correct *river* formula, when applied to observations made upon water flowing with perfect uniformity, ought to give too *small* a mean velocity." It appears to follow that such a formula, applied to a shallow, turbulent loaded stream, would give a velocity far in excess of the actual mean velocity. In other words, measured mean velocity in a natural stream is always less than the theoretical velocity because of the obstructions to free flow.

portional to the square root of the kinetic energy, ($V = \sqrt{E/\frac{1}{2}m}$), and only under special conditions may it approach such a value; in many cases it may be so small as to be proportional to the third or fourth root of the kinetic energy, or even much smaller. An extreme case is the complete loss of velocity at the end of intermittent desert streams, usually attributed to loss of volume, decrease in gradient and hydraulic radius, and excessive overloading.

The energy effective in moving a load, of course, is kinetic; and this energy varies as the square of the velocity. The greatest velocity attainable under natural conditions is that of free fall; anything less involves a loss in energy. Anything which promotes free-flow velocity of a stream increases its capacity to do work; and "each element of load, by drawing on the supply of energy, reduces velocity."[22] Consequently, the measured mean velocity of a loaded stream may bear no definite relation to the effective velocity which may be computed for that stream, and it may bear no predictable relation to either the capacity or the competence of the stream. Gilbert said that "capacity for transportation is not statable in units of energy," and consequently there seems no logical analysis by which it can be estimated.

The whole attempt to relate capacity (as defined by Gilbert) to the measured velocity of a stream or to the discharge of a trough is fraught with frustrations. In the first place, the velocity measured represents not the working energy but the wasted energy. It is like measuring the velocity of the waste waters below a turbine to compute the water power of the whole installation. It is putting the cart before the horse.

[22] Gilbert, p. 11 of ftn. 2 (1914).

FORMULAS APPLICABLE TO VELOCITY, CAPACITY, AND COMPETENCE

The safest and sanest formula for teachers of geology to apply to the behavior of a stream relates to the variation in velocity, upon which all work depends. This is the classic old Chezy formula,[23] $V = B\sqrt{rs}$, in which V equals velocity of the unencumbered stream, B is a constant, r is the hydraulic radius, and s is the slope. Obviously, velocity cannot vary without a change in other factors, all of which influence transportation. For the sake of analysis, we have assumed the possibility of variations in velocity, with all other factors being constant, contrary to actual conditions.

Actual conditions, free from wide variations in factors other than velocity, gave rise to the formula of C. Lechalas,[24] based upon sand movements in French rivers with sandy bottoms, the capacity of which varied as the square of the mean velocity with two empirical constants, both of which depended upon local conditions,

$$C = K(V_m^2 - k)\ .$$

For competence we have available the formula of John S. Owens,[25] derived from particles shifted by tidal currents, under particular, local (and rather restricted), natural conditions,

$$D = \frac{0.059}{G - 1} V^2\ ,$$

where D is the diameter of the pebbles in feet, G the density of the pebbles, and V

[23] Hydraulic engineers much prefer Manning's formula. It states that

$$V = \frac{1.486}{n} r^{2/3} s^{1/2}\ ,$$

in which V, r, and s are used as above and n is a roughness factor.

[24] "Note sur les rivières à fond de sable," quoted by Gilbert, p. 195 of ftn. 2 (1914).

[25] Quoted by Gilbert, *ibid.*, p. 163.

the speed of the current in feet per second. By choosing rock with specific gravity of about 2.8 and expressing the diameter (d) of the pebbles in inches, we may greatly simplify the formula by writing

$$d = 0.04V^2 .$$

This indicates that the diameter of the pieces rolled varied as the square of the velocity and that the volume of the pieces varied roughly as the sixth power of the measured velocity. This appears to confirm the usually accepted statement, unless the measured velocity in this case was actually close to the effective velocity. Actually, the latter was the case because the load moved was small, in proportion to the discharge, and because the range in velocity changes was not large. In discussing his own experiments it was recognized by Gilbert[26] that the effective velocity must be the bottom velocity, which he found so difficult to measure that he had to compromise by taking the so-called "mean" velocity. Practically, the effective velocity cannot be measured directly.

It would be convenient, if it were possible logically, to relate the capacity and the competence of a stream merely to its measured velocity; but the measured velocity is not the effective velocity and bears no simple relation to it. Quantitative experiments in flumes must be so rigidly controlled that they bear little direct application to natural streams. Since the slope and hydraulic radius of a stream vary from place to place along the course at any time and since the hydraulic radius and slope change at any place from time to time with changes in volume, it is the control by slope and hydraulic radius which gives us a quantitative value for the effective velocity. These are factors which are measured in hydrographic surveys and are available to students of stream action. One is able, therefore, in any special case to relate measured velocity with the computed value of free-flow velocity and to observe the amount of load and the maximum size of particles shifted, as Lechalas and Owens, quoted above, have done. But observations of current velocity alone cannot be applied properly to any form of stream behavior without regard to other controlling factors—especially quantity and fineness of load moved, slope, and hydraulic radius, which always change in natural streams when velocity changes. To overemphasize the implications of measured velocity is grossly misleading.

[26] *Ibid.*, p. 10.

CONCLUSION

The writer notes that many authors have confused competence and capacity of streams; some have been completely misled as to the relations of capacity and competence with stream velocity; and others are applying incorrectly to stream flow the empirical formulas based upon flume experiments.

It is thought that the velocity of free flow in a stream ought to be distinguished consistently from the residual velocity of a stream which has already used much of its energy in geological work. Empirical formulas based upon waste discharge or waste-water velocities in flume experiments should not be applied without specific reservations to the behavior of natural streams. It is recommended that velocity be discussed in geology textbooks according to Chezy's formula, with due emphasis upon the effects of loading, slope, and hydraulic radius of the stream.

It is suggested that "competence" should be defined in terms of force. Com-

petence may be related to the sixth power of the effective velocity; and in streams whose discharge is large in proportion to the load carried, competence varies very nearly as the sixth power of the measured velocity.

"Capacity," as defined and consistently treated by Gilbert, varies widely in reference to measured mean velocity because of concurrent changes in slope, hydraulic radius, and quantity of load carried; and it is not statable in terms of energy. Consequently, it is suggested that "capacity," as so defined, be omitted from textbooks of general geology and that the potential energy of a natural stream be described in terms of mass and effective velocity. Similarly, one may refer to the potential power of a stream which may be applied to geological processes, leading on to the actual power which might be generated by suitable "harnessing" of the stream's potential energy.

ACKNOWLEDGMENTS.—The writer has been encouraged to publish this study by many of his colleagues, both at the University of Illinois and elsewhere. Fortunately, an early draft of this article was submitted to Dr. M. King Hubbert, who made several trenchant and highly constructive criticisms. The writer has taken advantage of these various helps but reserves for himself the onus for errors which remain, for the general idea advanced, and for criticism of numerous authors. If the views or statements of these authors have been interpreted incorrectly, it has been done without intention and certainly without malice. It must be evident that this study would have no point were it not for the quotations from reputable authors. As a matter of fact, any credit or service which may accrue from this study may be shared by those who gave occasion for it—the authors who have been quoted.

STYLOLITES WITH FILMS OF COAL

PARIS B. STOCKDALE
University of Tennessee, Knoxville

ABSTRACT

Recently discovered stylolites in quartzite, associated with thin films of coal, reveal significant evidence bearing upon the theory of origin. The relationships between the stylolite-seams with the coal partings and the sedimentary structures of the bedrock give conclusive support to the solution (secondary) theory of origin, as opposed by the pressure (primary) theory.

INTRODUCTION

The rare occurrence of stylolites in rocks other than such carbonate types as limestone, dolomite, and marble has been recognized for several years.[1] A recent discovery of a type of stylolite heretofore undescribed deserves brief mention not only because of its peculiarity but because of its bearing on the mode of origin of stylolitic phenomena.

The stylolites under consideration were observed in a 2-inch drill core taken from quartzitic sandstone of Pottsville age (probably the Gizzard formation) on Lookout Mountain, near Mentone, DeKalb County, Alabama.[2] The stylolites are peculiar because of the thin, black, coal films associated with them, as well as because of their occurrence in quartzitic sandstone. They are significant because of their structural relations, which contribute evidence in support of the solution theory of origin.

OCCURRENCE AND DESCRIPTION

The rock is a light-gray, medium-grained, very hard, compact quartzitic sandstone. It is virtually a quartzite. Stylolites were previously reported from similar rock of approximately the same geologic age along the Cumberland Escarpment in eastern Tennessee.[3] The latter, however, did not carry coal films. The rock of the drill core under study contains, in addition to the quartz which predominates, scattered miniature chunks and flakes of coal. The thin films of coal sharply demarcate the stylolite-seams. The coal is highly metamorphosed, is of very low volatility, and does not ignite freely. No true graphite, however, was observed.

The rock is laminated rather distinctly. This feature is accentuated by the occasional concentration of carbonaceous material, giving rise to darker streaks, alternating with the lighter ones. The laminae lie at an angle of 25° from the plane perpendicular to the axis of the drill core, suggesting a 25° angle of dip of the strata.

The stylolites are small. Individual columns range from almost microscopic size to not over $\frac{1}{4}$ inch in length. Some of the columns have nearly parallel sides and blunt, rounded ends. Miniature striations occur on some of the sides. Dominantly, however, the stylolites are tapered and pointed—the *Drucksuturen* type of the German writers (see Figs. 1 and 2).

[1] W. A. Tarr, "Stylolites in Quartzite," *Science*, Vol. XLIII (1916), pp. 819–20; Paul H. Price, "Stylolites in Sandstone," *Jour. Geol.*, Vol. XLII (1934), pp. 188–92; Paris B. Stockdale, "Rare Stylolites," *Amer. Jour. Sci.*, Vol. XXXII (5th ser., 1936), pp. 129–33.

[2] United States Bureau of Mines, Project 810, "Low-Volatile Coking Coal on Lookout Mountain," Don M. Coulter, project engineer; under the direction of Richard W. Smith, district engineer of the Fifth District, Eastern Region.

[3] Stockdale, pp. 130–31 of ftn. 1 (1936).

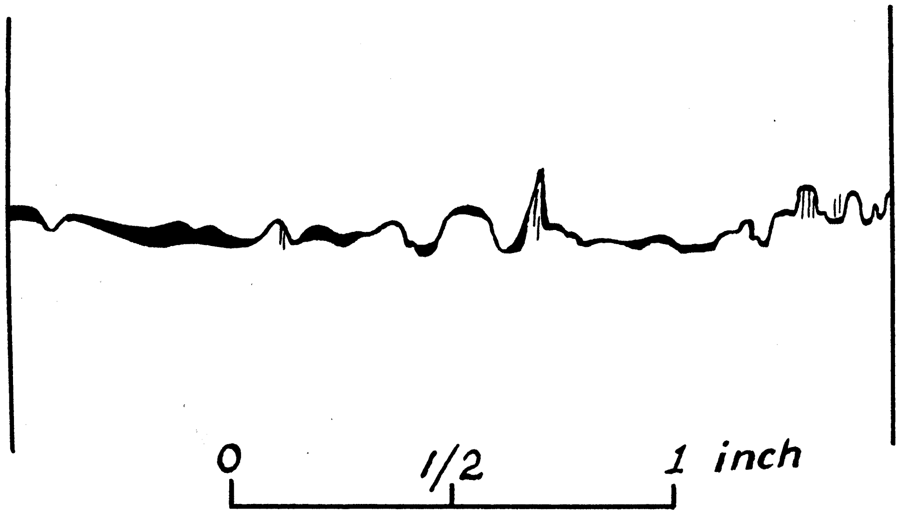


FIG. 1.—Sketch of a stylolite-seam showing the coal parting and the variations in the size and shape of individual stylolite columns.

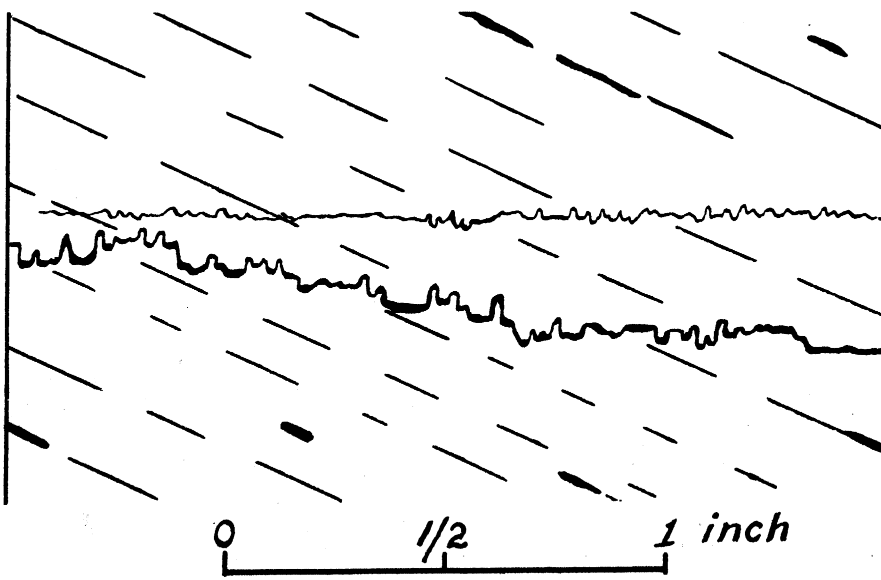


FIG. 2.—Sketch of two small stylolite-seams showing their structural relationships with the original, diagonal lamination of the bedrock.

The coal films associated with the stylolite-seams are the significant feature. Although thin in actual measure, the coal films are comparatively thick in relation to the length of the interpenetrating stylolite columns (Fig. 1). Stylolite surfaces which are revealed by separating the rock along stylolite-seams are rough and pitted. They are typical solution surfaces. The incrusting coal films can be readily scraped off and examined.

STRUCTURAL RELATIONSHIPS AND THEIR BEARING ON THE ORIGIN OF THE STYLOLITES

Stylolite-seams in the drill core of this study have a position which was approximately horizontal in the bedrock, cutting through and across the inclined laminae at an angle of approximately 25°. They do not follow lamination planes. They were developed along transverse fractures across the tilted bedding. The associated coal films, too, cut through and across the original sedimentary layers (Fig. 2).

Obviously, such a structural relationship has a significant bearing on the time and mode of origin of the stylolites. Instead of being developed as primary structures along normal planes of sedimentation at or near the time of deposition of the original sediment (sand), the stylolite-seams came into being as secondary features subsequent to the original deposition and compaction of the sediment. They were developed along fracture-partings which originated at a time not only later than the original consolidation of the sediment into sandstone but even later than the metamorphism into the quartzite. Thus there is revealed another finding to be added to the already existing preponderance of evidence in support of the belief that stylolites are of secondary origin, as held by the solution theory.[4]

The solution theory contends that the stylolites result from differential chemical solution in hardened rock, under some pressure, on the two sides of a parting of some sort (such as a bedding plane, lamination plane, suture, or crevice), the individual portions of the one side fitting into the dissolved-out parts of the opposite, the interfitting taking place slowly and gradually as solution continues. According to this idea, stylolites are a strictly secondary phenomenon, developed after consolidation and hardening of rock material—a feature that may be in the making in bedrock today under proper conditions. The films of coal associated with the stylolites are a residual concentrate—the insoluble constituents left from the dissolving of the stone. The originally disseminated bits of carbonaceous material became concentrated along the fracture where solution occurred.

It is true that most stylolites, and particularly large stylolites, are found in carbonate rocks, such as limestone, dolomite, and marble. These are the types in which one would expect most frequent and abundant solution by circulating ground waters. However, the occurrence of stylolites in rocks other than those which are readily soluble, such as quartzite, does not disprove the solution theory. One must bear in mind the fact that instances of such are comparatively rare and that the structures are comparatively small and are not strongly developed. Although solution of quartz in normal ground waters may be impercep-

[4] Stockdale, "Stylolites: Their Nature and Origin," *Ind. Univ. Study No. 55* (1922), pp. 1–97; "Stylolites: Primary or Secondary?" *Jour. Sed. Pet.*, Vol. XIII (1943), pp. 3–12.

tible, as measured by laboratory experiments and observations covering but few years of time, dissolving which has gone on through the vast time allowed by geology should occasionally be discernible. That even quartz is slightly soluble is recognized by present-day chemists.

By supporting the idea of secondary origin of stylolites, the structural relationships described above preclude the notion of primary origin, as contended by proponents of the "pressure" theory. According to the latter theory, stylolites originate while sediments are in an unconsolidated state, due to differential pressure and compaction. The theory fails to explain such a phenomenon as the coal films described above, which lie in a position which could not possibly have been one of original sedimentation.

SOLUTION OF LIMESTONE BENEATH HALES BAR DAM

JOHN W. FRINK
Tennessee Valley Authority, Chattanooga, Tennessee

The purpose of this paper is to point out an example of rapid solution of limestone in the foundation of Hales Bar Dam, built on the Tennessee River in 1905–13 by a private power company. The foundation consists of nearly flat-lying, massive Bangor limestone (Mississippian), averaging better than 95 per cent calcium carbonate. The cavernous and weathered condition of the rock caused considerable trouble in securing the foundation at the time of construction and resulted in water passing beneath the dam as soon as the reservoir was filled. Various programs designed to stop this leakage were attempted during the 26 years between completion of the structure and its acquisition by the Tennessee Valley Authority. Although the total cost of these programs amounted to several million dollars, none was successful.

In 1940 the Authority started a program of grouting, combined with a cutoff wall to stop the leakage under the dam. All visible signs of leakage from the thirteen boils in the tail water have now been eliminated, and the project is virtually completed. The cutoff wall was made by drilling 18-inch diameter holes through the rock along the upstream face of the dam. First, a row of holes, spaced on 2-foot centers, was drilled, lined with asbestos-cement pipe, and filled with concrete. Interconnecting holes were then drilled slightly upstream, biting into the adjacent, concrete-filled holes. When the latter holes were lined and concreted, a continuous cutoff wall resulted (Fig. 1).[1]

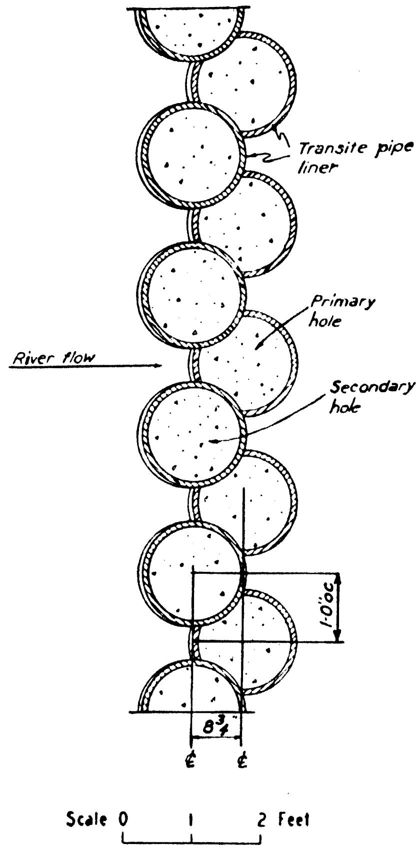


FIG. 1.—Plan of 18-inch holes (after Schmidt)

In one part of the foundation three of the secondary or intersecting holes were

[1] L. A. Schmidt, Jr., "Flowing Water in Underground Channels, Hales Bar Dam, Tennessee," *Proc. Amer. Soc. Civil Engineers*, Vol. LXXIX, No. 9 (1943), pp. 1417–46.

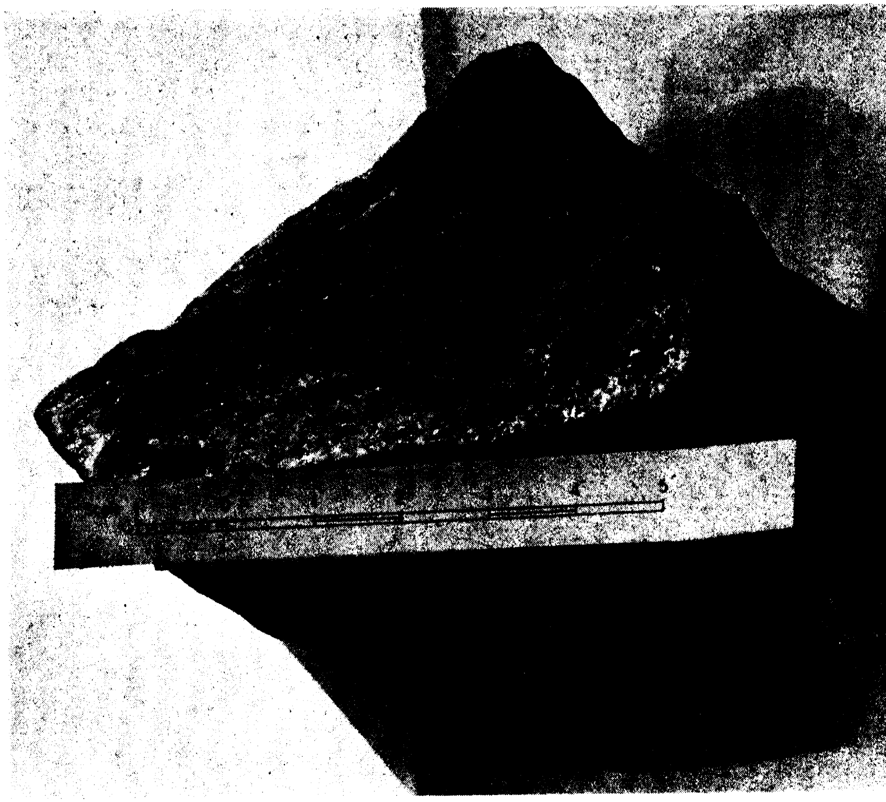


FIG. 2.—Core from secondary 18-inch hole, showing surface cut by a primary hole and later reduced by solution.

FIG. 3.—Same sample as Fig. 2, showing curvature of cut surface.

drilled more than a year after the primary holes. When the core which was recovered in the former holes was studied, it was noticed that the rock surface which had been cut by the drilling of the primary holes was pitted and in places reduced by solution as much as 0.1–0.13 inch (Figs. 2 and 3). Solution in all cases occurred below a cavity from 1 to 3 feet thick, about 60 feet beneath the surface. Moderate velocity was observed in this cavity, indicating active circulation, which brought quantities of water in contact with the rock surface.

The amount of reduction was determined in the following manner: when a wooden template, cut to fit the outside

diameter of an 18-inch hole, was placed on the curved surface of the core, it was found that the smooth area in the lower left of Figure 2 had not been reduced at all and that the shale parting forming the ridge in the center was not reduced appreciably. The depth from this ridge to the flanking pits proved to be a maximum of 0.13 inch, but 0.1 inch is the more general rule.

TABLE 1

SOLUTION OF LIMESTONE IN 18-INCH CORES

Hole Number	Date Drilled	Hole Intersected	Date Drilled	Interval in Months	Solution
C 498.......	9-23-42	C 422	6-30-41	15	0.1–0.13-inch reduction of limestone; shale partings not reduced
C 492A.....	10-24-42	C 429	9-10-41	13	Slightly pitted; no reduction aside from pits
C 499.......	11-12-42	C 422	6-30-41	17	Surface pitted; reduction slight

Portland P. Fox[2] has demonstrated solution of boulders in the Tennessee River at Watts Bar Dam to amount to 0.5 inch in 25 years. In comparison, the rock at Hales Bar would be reduced about 2.5 inches in the same length of time. In comparing the two rates, one must consider the following:

1. The Bangor limestone at Hales Bar, one of the most soluble limestones in this region, is considerably more soluble than the Nolichucky limestone boulders at Watts Bar.

2. The boulders described by Fox were above water-level during the low-water stages of the river. Thus they were free from solvent action a part of each year.

The rate of solution described here is not comparable to that under normal ground-water conditions, but it indicates the rate of solution of a very soluble limestone under conditions of free water circulation.

[2] "Rate of Solution of Some Limestone Boulders in the Tennessee River," *Jour. Tenn. Acad. Sci.*, Vol. XVI, No. 4 (1941), pp. 333–35.

REVIEWS

Index Fossils of North America. By H. W. SHIMER and R. R. SHROCK. New York: John Wiley & Sons, Inc., 1944. Pp. 837; pls. 303; figs. 5. $20.

The authors state the primary purpose of *Index Fossils of North America* to be "to describe and illustrate as many as possible of those fossils which can be used to identify and date formations and to correlate them from one area to another." To this end, genera which have a short vertical range and a wide geographical distribution have been considered to be most important, but genera which have been reported from localities distant from each other have been included because they must eventually be found in intermediate positions; genera with long vertical ranges are included because they are common; recently described genera and species which may prove useful are cited; and some genera are included "merely to round out the general picture of invertebrate fossils." Thus it can be seen that this book contains much more material than one would expect from its title.

Index fossils, as considered by Shimer and Shrock, consist of genera, form genera, species, form species, and varieties. A form genus in this work includes (1) "a series of related genera which have resulted from the splitting up of an old familiar genus," (2) "a complex of genera which are known to be distinct but which have not as yet been named because all have the same general habit," and (3) "the residium after a genus has been partly revised." It should be noted that the authors do not expect paleontologists to make rapid and final identifications of material merely by thumbing through this book, for they state: "Judicial use of the illustrations should help the paleontological investigator toward identifying and dating his fossils; but if precise determinations are required, an adequate library will be necessary and the present work can then only serve as a guide for further research."

The principal portion of the book is arranged in the usual taxonomic order by phyla. A concise discussion of the morphology of each class or other large taxonomic group precedes the discussion of the genera and species; along with this discussion is a general list of the more appropriate references, carried as a long footnote. In general, genera are arranged in taxonomic order, but species are usually arranged in reverse stratigraphic order. Common generic synonyms may be cited, and synonyms of species occasionally are cited. A typical reference to a genus and its species takes the following form. The recognized generic name with its author and date of proposal is given a center heading in capitals and may be followed by citations of common synonyms in parentheses. The name of the genotype as originally proposed, in italics, and marked by an asterisk, together with its author's name and the plate and figure numbers (if illustrated in this book), all set in parentheses, begins a paragraph of generic characters in telegraphic style. At the end of the paragraph is the geologic age of the fossil in bold-face type, and the localities, generally by states, in light-face type. Each species cited starts a new line below the above paragraph, with the binominal in bold-face type and the name of the author in light-face type, followed by the plate and figure numbers in parentheses. Any distinguishing features are next noted, and the age and geographic distribution are given as with the genus. Plate designations throughout are in bold-face type, separated from the figure numbers in light-face type by a dash.

As most paleontologists already realize, the authors have received extensive co-operation from their colleagues in the science, all of which has been scrupulously acknowledged. Some sections were prepared *in toto* by other authors, and some sections were revised by other authors.

Perhaps the best way to appreciate the wide coverage of the book is to examine the work quantitatively. The following figures in Table 1 were arrived at by one count of the species, genera, and figured species; actually the total figure here given is somewhat less than the actual figure because genotype species were not counted unless they were listed also as index fossils. Names in the column to the left are those in common use among paleontologists, rather than of each major taxonomic

group in the text, because it seems that a clearer idea of coverage can be obtained by referring to well-known groups. Separate phyla with subtotals are shown in order to condense the gross relationships. The first column shows the number of genera listed, the second shows "species" as counted, and the third shows relative proportion of species in each phylum or major group.

It can be seen that the greatest prominence is given to gastropods, brachiopods, crinoids, pelecypods, trilobites, cephalopods, ostracodes, and corals, in that order, each of which has more than two hundred species listed.

Practically every species mentioned in the text is illustrated, there being 4,543 species figured by the writer's count. There are said to be over 9,400 illustrations, in which case there is an average of two figures for every species. In order to accomplish this profuse illustration, the 303 plates are filled almost to the trimming edge with figures, and the general plan affords a plate facing nearly every page of text. The plates are quite pleasing in appearance and are remarkably uniform in tone, considering the number of different sources of figures. Plate explanations are set in very small type at the bottom of each page opposite a plate, and the binominals are set in bold-face type to facilitate locating them in this small type.

Some genera are represented only by the genotype and a large number are represented by only one other species. On the other hand, according to the index, 58 genera are represented by more than 10 species each, and 18 genera are represented by more than 15 species each. These latter are: *Ammonites*, *Arabellites*, *Atrypa*, *Bellerephon*, *Cardium*, *Chonetes*, *Goniatites*, *Hyolithes*, *Nautilus*, *Nucula*, *Orthis*, *Orthoceras*, *Ostraea*, *Pecten*, *Pentremites*, *Spirifer*, *Straparolus*, and *Turritella*. Of these, *Pecten* has 44 species listed under it, *Ammonites* 37, *Orthis* 33, and *Ostraea* 30.

In a work of this magnitude, which has not only taken years to compile but has involved the efforts of numerous collaborators, it is to be expected that inconsistencies will be present; therefore, one is not surprised to find variations in the style and makeup of the groups. One inconsistency that affects the use of the book is the variation in listing of species, which are generally placed on separate lines after the generic diagnoses, but here and there occupy entire paragraphs and are less easy to scan

TABLE 1

Group	No. of Genera		No. of Species		Percentage of Species in Group
Protozoa		126		207	4.5
silicoflagellates	8		8		
foraminifers	98		156		
fusulinids	10		33		
radiolarians	10		10		
Porifera		37		52	1.1
sponges	37		52		
Coelenterata		165		349	7.6
stromatoporoids	13		23		
graptolites	31		84		
conularids	4		9		
corals	117		233		
Echinodermata		336		704	15.3
cystoids	22		27		
edrioasteroids	11		13		
blastoids	11		29		
crinoids	242		549		
stelleroids	11		11		
echinoids	36		69		
holothurians	3		6		
Vermes		15		92	2.0
worms	15		92		
		72		88	1.9
? conodonts	72		88		
Bryozoa		135		193	4.2
bryozoans	135		193		
Brachiopoda		305		578	12.6
brachiopods	305		578		
Mollusca		734		1,550	33.8
pelecypods	187		511		
gastropods	279		663		
pteropods	4		4		
scaphopods	2		17		
chitons	7		8		
cephalopods	247		319		
other molluscs	8		28		
Arthropoda		420		753	16.4
agnostids	13		15		
trilobites	219		352		
ostracodes	124		293		
other arthropods	64		93		
Plantae		14		17	0.4
charaphytes	8		9		
algae	6		8		
Probable organisms	6	6	7	7	0.2
Total	2,365		4,590		100.0

than when listed normally. Another is the frequent omission of differentiating characters among species of a genus. Apparently it was felt that the illustrations would serve the same purpose, but not all species listed are illustrated. As a matter of fact, many species are figured but occur in the text without discussion, and a few figured species are not mentioned in the text at all. Finally, most groups are arranged in biologic order, but some groups are arranged in stratigraphic order because specialists were interested in fossils of certain ages, and other groups are arranged in alphabetical order because the existing taxonomy of higher orders was considered to be unsatisfactory.

Those who are expecting this book to be merely a revision of *North American Index Fossils* will be surprised, for there is little to remind one of the earlier work. Text material has been enormously increased and the length of descriptions shortened accordingly. Illustrations are likewise more numerous and are contained on plates rather than being distributed as line cuts in the text. Finally, there is no section devoted to stratigraphy.

Index Fossils of North America is sure to be a welcome handbook for paleontologists, for it far exceeds one's reasonable expectations. For some reason, paleontologists have been loath to cite names in synonymies from the earlier work by Grabau and Shimer, but this duty scarcely can be escaped now, for not only is the volume replete with new binominal combinations but new names are introduced here and there. This book will not only be widely used by laboratory paleontologists, as well as field stratigraphers, but is admirably suited to the needs of teachers of advanced paleontology. The authors have produced a truly monumental work that should be on every paleontologist's bookshelf.

WILLIAM H. EASTON

Lake and Rastall's Textbook of Geology. Revised by R. H. RASTALL. London: Edward Arnold & Co., 1941 (reprinted, 1943). Pp. viii+491; pls. 32; figs. 128. 25*s*.

Dr. Rastall's revision of the text of the well-known "Lake and Rastall" (first published in 1910) has been so thorough that he has made a new book of it. As the author points out, some sections have had to be shortened in order to make room for new matter, because, in his opinion, "in course of time parts of the subject have become of greater and others of less importance." He has taken particular care in this way to avoid "the common error of making successive editions of a textbook more and more complicated, so that eventually the whole founders under a mass of unintelligible detail." No doubt because of considerations of war economy all the photographic illustrations are the same and in the same order and described in the same words as in the first edition; but the quality of the photographs themselves and of the half-tone blocks would alone warrant their retention. (One notes that the misprint "nick" for "neck" in the caption of Pl. XXVIII [1], which appeared in the first edition, has not been corrected.)

The very inferior paper used in the 1943 impression does scant justice to the excellent photographs, as a comparison with earlier editions will show. The new volume, though the number of pages in it is approximately the same, is of little more than half the thickness of the first edition.

Perhaps an unjustifiable lack of faith in the continued popularity of this textbook has led to a decision to print only a small number of copies. This, at any rate, may be the explanation of the very high price at which the new edition is listed. It is a matter for regret that a book potentially so useful for beginners should be so expensive as necessarily to restrict its circulation.

Although the treatment is definitely for beginners, the author has inserted (as footnotes) a few well-selected references to sources of information. He has, indeed, struck a happy mean between too few and too many of these. The purpose of including references—whether as footnotes or in a more pretentious bibliography—is twofold: (1) to give credit where credit is due and at the same time shift some responsibility from the author's shoulders and (2) to afford to students an opportunity to extend their supplementary reading. It is usual to omit those in the first category from books that are purely students' textbooks; but the lists provided in many books for the avowed purpose of encouraging undergraduate reading are so voluminous as to defeat their own purpose. In the volume under review the selection is judicious; but the list might have been expanded somewhat with advantage along certain lines.

A general textbook of geology is so encyclopedic in its scope that it seems invidious to pick out any section for special comment and criti-

cism. One must bear in mind that the need for conservation of space has constantly curbed free expression. Condensation of matter (consistently with clearness), with the omission of all items of information that are either irrelevant to the immediate topic or unnecessary for the beginner, must be the keynote throughout. In this book the balance between diverse topics is well maintained; and, while a surprising amount of sound information is imparted on every page, the author has found it possible to avoid stodginess and to maintain an easy, breezy style.

The separate chapter on metamorphism, found so readable and so helpful by many undergraduates, which was a feature of the earlier editions, has unfortunately been squeezed out; but another useful chapter on ore deposits still remains.

As far as page 299 the treatment is general; the book is of world-wide scope. The stratigraphical part, however, pages 300–477, treats exclusively of the geology of the British Isles. It would be out of place, perhaps, to criticize this adversely, considering the scope of the book and the fact that it is written primarily for students in those isles, but from the point of view of these very British students it seems unfair to convey the impression—as the book certainly does—that British stratigraphy, notably that of the Devonian, Permian, Triassic, and Tertiary, is an epitome of world geology. For more advanced students in America and in the British overseas dominions, whose stratigraphic fodder consists primarily of their own local geology, the up-to-date summary account of the stratigraphy of Britain here provided will be found to be well worth reading. To them it will appear appropriate that for purposes of description of the local geology of the British Isles the names "Permian" and "Triassic" are dropped in favor of a revival of "New Red Sandstone," though it may seem odd that British undergraduates are to be kept in ignorance of the important marine facies of these systems in foreign lands.

C. A. COTTON

Elements of Mineralogy. By ALEXANDER N. WINCHELL. New York: Prentice-Hall, Inc., 1942. Pp. xiii+535; 1 colored plate; figs. 468. $5.00.

The subject is treated in six parts: crystallography and physical, chemical, descriptive, economic, and determinative mineralogy. Crystallography (120 pages) is mainly morphology (9 pages of X-ray work) as applied to models. There is no attempt to deal with the measurement, calculation, and projection of actual crystals, though the stereographic projection is mentioned and is used qualitatively to indicate distribution of symmetry and the face-poles of general forms. The author wisely employs Mauguin symmetry symbolism. Line drawings appear beside photographs of models; Miller indices (in place of letters) are used on the drawings. Fifteen classes are discussed in detail.

Physical mineralogy (56 pages) includes optical crystallography (32 pages) in abbreviated form. Two-thirds of chemical mineralogy (42 pages) is devoted to blowpipe analysis. Minerals are classified into ten divisions in descriptive mineralogy (176 pages), of which one (organic compounds) is not covered. This portion of the book is unique for an elementary text, in the sense that here a praiseworthy attempt is made to bring the modern science of mineralogy to the level of the college classroom. Many minerals are recognized members of systems showing properties variable with changes in chemical composition; and 63 diagrams, of the sort found in the author's *Elements of Optical Mineralogy*, are here introduced. Many of these, of course, represent simplifications of the facts in so far as we now understand them; most could be reduced in size advantageously. Probably this phase of the book is more advanced than is warranted by what is offered in the preceding portions.

The section on economic mineralogy (32 pages) omits silicates and covers other minerals in terms of basic and acidic divisions, with elements arranged in the order of the periodic table. Peculiarly, groups V B and VI B are put with the acids; and groups III A and IV A are split between the two divisions. Alphabetical order, in terms of metals and nonmetals, would be an improvement.

Determinative mineralogy (77 pages) contains four tables, based on (1) streak and specific gravity, with four main subdivisions; (2) hardness, streak, and gravity, with seven main subdivisions; (3) cleavage, luster, and hardness, with five main subdivisions; and (4) refractive index. These tables are a valuable addition to the book. They and the mineral diagrams constitute the oustanding features of the work. They are followed by a Glossary and Index.

D. J. F.

Geology and Ground-Water Resources of the Island of Maui, Hawaii. By HAROLD T. STEARNS and GORDON A. MACDONALD. Territory of Hawaii, Division of Hydrography, Bull. 7. Prepared in co-operation with the U.S. Geological Survey. Honolulu, 1942. Pp. xiv+344; pls. 44; figs. 46.

Maui, the second largest island in the Hawaiian group, was built by two volcanoes—East Maui, or Haleakala, and the West Maui Volcano, now dissected into several high peaks. A flattish isthmus connecting them was formed chiefly by lava flows from Haleakala. The oldest rocks exposed are very permeable basalts which were extruded probably in Pliocene and early Pleistocene time from several rift zones. Beginning with these outpourings, the development of the island has been worked out in thoroughgoing fashion, well described and admirably illustrated, stage by stage, by diagrams and sketches of artistic merit.

Part I by Stearns, constituting the bulk of the report, presents the geomorphology, general geology, and ground-water resources, first of East Maui and then of West Maui. Particularly interesting is the so-called "crater" of Haleakala, once thought to be a vast caldera but now known to have resulted chiefly from stream erosion. The top of the mountain has been removed by the headward erosion of the Keanae and Kaupo valleys from opposite sides of the island. Owing to the nature of the rocks and the local conditions, these valleys grew headward as steep-walled amphitheaters which flared out in ground plan as they approached the higher middle portion of the island. Eventually they met, and the dividing wall became greatly reduced. Renewed volcanic activity produced several small cinder cones in the great summit amphitheater, while their flows partially filled the two valleys leading to the sea, thus contributing to the illusion of a caldera. This history, and much more, is beautifully pictured in a succession of colored plates representing eight stages in the development of Maui.

Part II is a special study of the Nahiku area in East Maui, while Part III gives the detailed petrography of the whole island. Both of these are by Macdonald.

This report contains a wealth of varied information, and its illustrations—airplane photographs and diagrammatic sketches alike—are unusually effective and pleasing to the eye.

R. T. C.

Geology. By JOHN A. ALLAN. (Research Council of Alberta Report No. 34.) Edmonton: University of Alberta, 1943. Pp. 196; pls. 45; maps in pocket. $1.50.

Annual reports on the geological work carried out by the Research Council of Alberta were published up to and including 1935, when they were discontinued. Report No. 34 now gives some of the results of a number of geological survey projects undertaken by J. A. Allan between 1936 and 1941 inclusive. In Part I the general geology of Alberta is discussed briefly and the major geological formations described. Part II treats of the extensive salt deposit at Waterways in the McMurray district; Part III is devoted to the geology of Alberta soils in the 28,000 square miles where detailed soil surveys have been made; Part IV describes a large relief model of Alberta; and Part V describes and discusses the coal areas.

Of especial interest is the thick bed of nearly pure salt which measures 199 and 211 feet in two excellent drill cores recently obtained. For 200 feet above the salt bed are many alternations of anhydrite and gypsum with some thin beds of dolomite and shale. Resting on this sequence are limestones and shales carrying an early Upper Devonian fauna, which may be correlated with the Upper Devonian lying immediately above the Silurian at Peace Point on the Peace River. Although no fossils were encountered in the evaporite series, it is considered Upper Silurian in age. Is this a correlative of the Salina of Michigan and New York indicating similar conditions in a long, narrow arm of the sea reaching across the continent from the Mackenzie arctic region?

Banding is apparent throughout much of the rock salt, and the marked regularity in the thickness of the salt layers seems to suggest annual banding. An 8-foot sample of the core was found to contain 142 distinct layers. It is, however, probably not yet safe to make chronological estimates on this basis.

R. T. C.

VOLUME LIII NUMBER 3

THE JOURNAL OF GEOLOGY

May 1945

MIDDLE ORDOVICIAN LIMESTONES FROM LEE COUNTY VIRGINIA TO CENTRAL KENTUCKY[1]

GEORGE GARRETT HUFFMAN
1334 Daytona Drive, Corpus Christi, Texas

ABSTRACT

Limestones previously mapped as "Lowville" in Lee County, Virginia, between the cherty "Lenoir" and "Trenton" have lithologic and faunal zones that correlate with units to the east in Virginia and north in Kentucky.

The lower 250–300 feet of "Lowville" is stratigraphically similar to the "Ottosee" of Rye Cove, Virginia. The succeeding 550 feet has a thin *Stromatocerium* zone 65 feet above the base, one of *Hesperorthis* 10 feet higher; the base of a 100-foot zone with *Cryptophragmus* and *Camarocladia* is 55 feet above. The *Cryptophgramus* zone is lower "Moccasin" at Rye Cove, Virginia, and upper Camp Nelson in Kentucky. The upper 150 feet (Eggleston mud-rock facies) is the equivalent of the Tyrone of Kentucky; the two thick metabentonites in the Eggleston are like those in the Tyrone.

The Eggleston is overlain conformably by 40 feet of crystalline limestone carrying a Curdsville fauna. The succeeding 60–70 feet of alternating shale and limestone grading upward into yellow-brown, calcareous shale has the fauna of the Logana ("Hermitage") of Kentucky. The overlying "Catheys-Cannon" seems to correspond to middle and upper Trenton of Kentucky.

Lower Trenton age of the Eggleston and Tyrone is indicated by their conformability with the overlying Curdsville, which is homotaxial with the Kirkfield (Hull). Correlation of the Oregon is uncertain. The *Cryptophragmus* beds correspond to the Benner of Pennsylvania, classed as Black River Pamelia and Lowville.

INTRODUCTION

Middle Ordovician limestones have been known to outcrop extensively in Lee County, extreme western Virginia,[2] but their stratigraphy has been discussed only superficially, no detailed study having been undertaken. The section in the center of the Jessamine dome in Kentucky, 150 miles northwest, although unexposed at the base, has been described more adequately. Lee County lies intermediate between the thick clastic section of the Middle Ordovician geosyncline in Virginia and the thinner limestone section in Kentucky; an understanding of its stratigraphy is therefore important. The writer determined the lithologic units in Lee County and recognized corresponding units in Kentucky. Subsurface studies previously showed northwestward thinning in the latter state;[3] the present work reveals the manner of convergence. More precise dating of the St. Peter sandstone is made possible. Subsequent to the completion of field

[1] A thesis submitted in partial fulfilment of the requirements for the degree of Doctor of Philosophy, in the Faculty of Pure Science, Columbia University (1944).

[2] Charles Butts, "Geologic Map of the Appalachian Valley in Virginia," *Va. Geol. Surv. Bull.* 42 (1933), pp. 1–50 with map.

[3] L. B. Freeman, "Present Status of St. Peter Problem in Kentucky," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXII (1939), pp. 1836–43.

study, similar rocks were described and classified in Tazewell County, Virginia,[4] 50 miles northeast of Lee County. Correlations can be carried between previously unrelated Appalachian and Kentucky sections. The regional stratigraphy is not adequately known, but progress is being made and the whole is gradually coming to form a co-ordinated picture.

The original plan was (1) to study and zone the Middle Ordovician rocks in Lee County, Virginia, and (2) to compare similar rocks in Kentucky. The need for suitable stratigraphic nomenclature for the Lee County section prompted investigation of regional stratigraphy of the Middle Ordovician of southwestern Virginia. The study results in a classification of the strata in Lee County (see Table 1). Since the field work was completed prior to publication of new nomenclature proposed for Tazewell County,[5] use of older terminology seems advisable; questionable formation names are placed in quotation marks. Inasmuch as study of the extension of the Tazewell units has been undertaken,[6] application of new stratigraphic names is withheld though comparisons are made.

STRATIGRAPHY OF THE LEE COUNTY SECTION

Regional Setting

Lee County lies near the outer edge of the folded Appalachian belt in the most northwestwardly thrust block where the sedimentary rocks form the broad, asymmetrical Powell Valley anticline extending from Tennessee into southwestern Virginia and plunging northeastward into Wise County. The Powell Valley anticline is part of the Cumberland thrust block,[7] a rectangular mass about 125 miles long and 25–30 miles wide, bounded on the northwest by the Pine Mountain thrust fault, on the southwest by the Jacksboro tear fault, on the northeast by the Russell tear fault, and on the southeast by a pair of parallel faults, the Hunter Valley thrust and the Wallen Valley thrust. These relations are shown in Figure 1. The Lee County sections are, for the most part, on the limbs of the Powell Valley anticline. The Glass Store or Hunter Gap section is south of the Wallen Valley fault; Rye Cove, Scott County, is south of the Hunter Valley thrust (Fig. 2).

The principal geosyncline of the Middle Ordovician seems to have been defined by a flexure about 25 miles southeast of the Powell Valley anticline near the present Saltville thrust. Thick clastic sediments (Blount Group) have been considered to disappear along this line, although recent studies indicate that they converge and grade into limestone at the flexure. The strata northwest of this line are in several thrust slices; beds deposited in farther separated areas have come into closer proximity. Attempts to account for the differences in lithologies and faunas of beds in adjacent blocks have resulted in several interpretations of the paleogeography and stratigraphy.

[4] B. N. Cooper and C. E. Prouty, "Stratigraphy of the Lower Middle Ordovician of Tazewell County, Virginia," *Bull. Geol. Soc. Amer.*, Vol. LIV (1943), pp. 819–86.

[5] *Ibid*, p. 884, Fig. 3.

[6] Prouty, "Middle Ordovician Limestones in the Median and Northwest Belts of Virginia and Tennessee" (abstr.), *Bull. Geol. Soc. Amer.*, Vol. LII (1941), pp. 1973–74.

[7] C. K. Wentworth, "Russell Fork Fault of Southwestern Virginia," *Jour. Geol.*, Vol. XXIX (1921), pp. 351–69; Charles Butts, "Fensters in the Cumberland Overthrust Block in Southwestern Virginia," *Va. Geol. Surv. Bull. 28* (1927), pp. 1–12; J. L. Rich, "Mechanics of Low-angle Overthrust Faulting as Illustrated by Cumberland Thrust-Block, Virginia, Kentucky, and Tennessee," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XVIII (1934), pp. 1584–96.

The Lee County sections lie wholly north of the Middle Ordovician geosyncline on the continental shield or craton zones from one thrust block to another, it is possible to establish correlations with sections nearer the geosyncline.

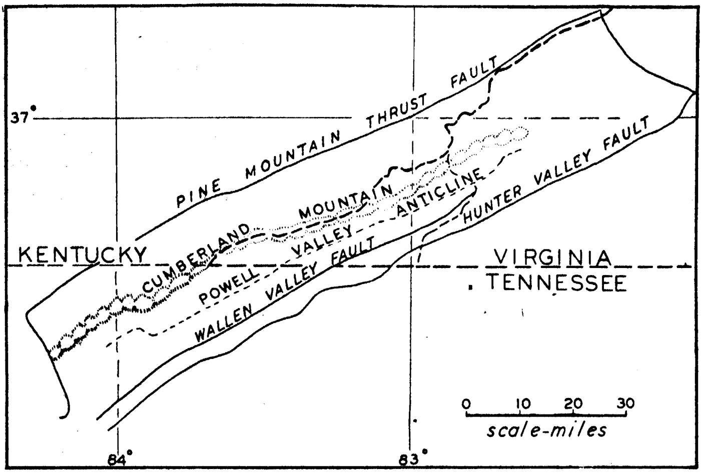


FIG. 1.—Cumberland overthrust block (after Rich, Butts, and others)

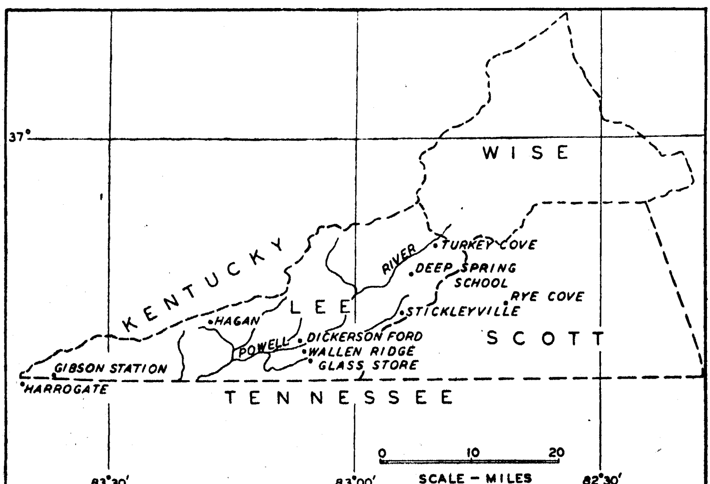


FIG. 2.—Index map showing location of sections in southwestern Virginia

in a position between the thick clastic sections southeast of the Saltville thrust and the central Kentucky section. By careful tracing of lithologic and faunal

EARLIER INTERPRETATIONS OF REGIONAL STRATIGRAPHY

In the southeastern belt of the Appalachian Valley, southeast of the Salt-

ville thrust, Butts[8] has proposed the following lithologic succession:

Lowville-Moccasin
Blount Group
  Ottosee limestone
  Hiatus: Tellico absent
  Athens shale
  Whitesburg limestone
  Holston limestone
Lenoir

The "Ottosee" was reportedly confined to the southeastern and median belts of the Appalachian Valley, the most northern outliers being at Rye Cove, Scott County, Virginia. The Athens shale did not extend northwest of Clinch Mountain or of the line of Clinch Mountain northwest of Burkes Garden, Tazewell County.

In the northwest belt of the Appalachian Valley, including Lee County, the rocks between the Beekmantown and the Cincinnatian (Reedsville) were mapped as (1) Stones River, (2) Lowville, and (3) Trenton.[9] The "Stones River" included the "Mosheim" and "Lenoir." The "Lowville" of Butts comprised 1,000 feet of dove-gray limestone and mud rock. Where the "Lowville" passes into red mud rock, the name "Moccasin" was applied. In areas transitional between the dominantly limestone facies and the dominantly mud-rock facies, Butts applied the term "Lowville-Moccasin." The overlying "Trenton Group" comprising the "Cannon" and "Catheys" limestones has been called middle and upper Trenton. The name "Eggleston" has been applied to a sequence of beds transitional between the Moccasin and Trenton; these beds are considered by some to be an upper shaly facies of the "Lowville-Moccasin."[10] Beds of lower Trenton age were considered absent in Wise County, east of Lee County;[11] however, R. E. Bates[12] and Butts[13] mentioned the possibility of lower Trenton strata in Lee County.

In Lee and Highland counties, in the northwest belt of the Appalachian Valley, the "Lowville" was said to rest on the "Lenoir" with an intervening hiatus due to the absence of the Blount Group.[14] Southeast of Clinch Mountain, the "Ottosee" reportedly rests on the Athens shale; in two belts northwest of Clinch Mountain, on the "Holston" limestone; farther northwest as at Rye Cove it succeeds the "Lenoir."[15]

Variations in the stratigraphic sequence in adjoining belts were attributed by earlier writers to parallel barriers separating the Appalachian geosyncline into several troughs of sedimentation,[16] a theory later modified by postulation of successive downwarpings from one trough to another.[17] P. E. Raymond[18] suggested that observed differences in adjoining belts might be due to distance from shore, depth of water, and other local

[8] "Geology of the Appalachian Valley in Virginia," *Va. Geol. Surv. Bull. 52* (1942), p. 148.

[9] Butts, pp. 1–50 of ftn. 1 (1933).

[10] R. R. Rosenkrans, "Stratigraphy of Ordovician Bentonite Beds in Southwestern Virginia," *Va. Geol. Surv. Bull. 46* (1936), pp. 85–112.

[11] G. W. Stose *et al.*, "Geology and Mineral Resources of Wise County and the Coal-bearing Portions of Scott County, Virginia," *Va. Geol. Surv. Bull 24* (1923), pp. 22–28.

[12] "Geology of Powell Valley in Northeastern Lee County, Virginia," *Va. Geol. Surv. Bull. 51B* (1939), pp. 37–94.

[13] Pp. 1–12 of ftn. 6 (1927).

[14] Butts, p. 178 of ftn. 7 (1942).

[15] *Ibid.*, pp. 170–71.

[16] E. O. Ulrich and Charles Schuchert, "Seas and Barriers of Eastern North America," *N.Y. State Mus. Bull. 52* (1902), pp. 633–62.

[17] Ulrich, "Revision of Paleozoic Systems," *Bull. Geol. Soc. Amer.*, Vol. XXII (1911), p. 292.

[18] "Middle Ordovician of Virginia and Tennessee" (abstr.), *Bull. Geol. Soc. Amer.*, Vol. XXXI (1920), p. 137.

conditions with possible continuous deposition from early Chazyan to late Trenton, except for slight emergence of short duration during and after Lowville.

Earlier interpretations of the Chazyan (Blount Group) have been challenged by Cooper and Prouty.[19] In Tazewell County and elsewhere in Virginia and Tennessee, they have divided the strata between the Beekmantown and Trenton into twenty-eight lithologic and faunal zones, constituting six formations. They state that many of the beds formerly assigned to the Blount group are of Stones River and Black River age; that "Mosheim" has been applied to two different limestones separated by 200–500 feet of beds; and the "Lenoir" of western Tazewell County is considerably older than "Lenoir" of the northwestern part of the county. Further, some of the beds called "Holston" overlie beds with a "Lowville" fauna. Cooper[20] has stated that the Athens formation is more extensive than previously believed; northwest of Clinch Mountain the Athens is represented by a limestone facies. In their final paper, which appeared after the writer had completed his field work, Cooper and Prouty[21] proposed a new system of classification for the Chazyan and Black River sequence in Tazewell County; they state that further use of the names Murfreesboro, Mosheim, Lenoir, Holston, Ottosee, Lowville, and Lowville-Moccasin, unless qualified, is inadvisable.

Stratigraphic Sequence

By way of introduction to a detailed discussion of the stratigraphy of Lee County, the major units which the writer has succeeded in distinguishing are summarized in Table 1. As previously indicated, traditional names are used; these will be found justified in the succeeding pages.

TABLE 1

STRATIGRAPHIC SUMMARY OF MIDDLE ORDOVICIAN OF LEE COUNTY

Description of Unit	Feet
"*Catheys-Cannon*": Blue, granular, even-bedded, fossiliferous limestone, thickness about	500
Hermitage: Calcareous shale (Zone 20) passing downward into gray, thin-bedded, fine- to medium-textured limestone and shale (Zones 19 and 18); abundant *Sinuites* and *Dalmanella*	60–70
Curdsville: Gray-blue, coarsely crystalline, thin to medium-bedded limestone (Zone 17) including at its base the "cuneiform" beds (Zone 16); characterized by *Sowerbyella curdsvillensis* and *Dinorthis pectinella*	40
Eggleston: Assemblage of yellow, calcareous mud rock and dove limestone (Zone 15) with a thin metabentonite at the base; two thick metabentonites 40 feet apart near the top	150
Moccasin: 135 feet of dove-gray, laminated limestone (Zone 14) underlain by a thick unit of yellow-green mud rock (Zone 13)	300
"*Lower Moccasin*": Dove-gray, sublithographic to medium-crystalline limestone with well-defined zones of *Stromatocerium*, *Hesperorthis*, *Camarocladia*, *Cryptophragmus* (Zones 7–12)	250
"*Ottosee*": Dove-gray to bluish-gray, locally argillaceous limestone with a bed of yellow, shaly limestone and red mud rock at the top and lenses of coarsely crystalline limestone near the middle (Zones 1–6)	300
"*Lenoir*": Dove-gray to brown, dense to finely crystalline, very cherty limestone	200
"*Mosheim*": Light-gray, sublithographic, thick-bedded limestone	25
"*Murfreesboro*": Sequence of basal conglomerate, silver-gray shale, red dolomite, and buff dolomite	85

[19] "Chazyan and Black River Stratigraphy in Tazewell County, Virginia" (abstr.), *Bull. Geol. Soc. Amer.*, Vol. LI (1940), p. 1924; Prouty, pp. 1973–74 of ftn. 5 (1941).

[20] "Athens Equivalents Northwest of Clinch Mountain in Southwestern Virginia" (abstr.), *Bull. Geol. Soc. Amer.*, Vol. LII (1941), p. 1893.

[21] Pp. 819–86 of ftn. 3 (1943).

This sequence is underlain by the Beekmantown dolomite and overlain by the Reedsville shale (Cincinnatian). Several exposures of "Stones River" are found in Lee County; hence a brief description of these units is included.

"STONES RIVER" OF PREVIOUS REPORTS

"MURFREESBORO" FORMATION

The Beekmantown dolomite is overlain unconformably by approximately 85 feet of red shale, red argillaceous dolomite, silver-gray shale, and buff dolomite. A prominent conglomerate containing angular fragments of chert and dolomite in a matrix of dolomite is at the base. This "Blackford" facies,[22] formerly included in the Beekmantown, is widely distributed in Lee County; good exposures can be seen along the railroad at Hagan, Virginia. The basal conglomerate shows in the field south of the highway east of Gibson Station, Virginia. There are additional exposures west of Arthur, Tennessee.

A section of a second facies, the "St. Clair" along Yellow Branch Creek, $5\frac{1}{2}$ miles southeast of Rose Hill has been described by Butts.[23] The relationship of the two facies has not been deduced from a study of the Lee County sections.

"MOSHEIM" FORMATION

The "Mosheim" limestone conformably overlies the "Murfreesboro" of Butts. The "Mosheim" is a light-gray, thick-bedded, sublithographic limestone (calcilutite) which breaks with a conchoidal fracture and weathers to rounded or fluted masses with a light-gray surface. About 25 feet is present in Lee County.

[22] Butts, p. 126 of ftn. 7 (1942).

[23] *Ibid.*, pp. 120–21.

"LENOIR" FORMATION

The "Lenoir," succeeding the "Mosheim," is a dove-gray to brown, dense to finely crystalline, thick-bedded limestone with large nodules of black chert which are especially abundant in the upper part. Good fossils are scarce, but remains of *Tetradium* are common on weathered surfaces. *Girvanella*, *Maclurites*, and *Gonioceras* also occur. The "Lenoir" is 150–200 feet thick in Lee County; it thickens southwestward into Tennessee, where as much as 150 feet of dove-gray, thin-bedded, somewhat shaly limestone intervenes between the "Mosheim" and the 95 feet of cherty upper "Lenoir" near Arthur, Tennessee.

"LOWVILLE" AND "TRENTON" OF PREVIOUS REPORTS

INTRODUCTION

The strata between the cherty "Lenoir" of Butts and the "Catheys-Cannon" limestone in Lee County can be divided into at least twenty lithologic and faunal zones (see Fig. 3). Zones 1 through 15 were included in the "Lowville" of Butts, so classified on the presence of *Cryptophragmus antiquatus* Raymond and *Tetradium cellulosum* (Hall), which were considered the invariable guide fossils of the Lowville from Ontario to Alabama. *C. antiquatus* is now known to occur throughout the Pamelia limestone of New York and Ontario while *T. cellulosum* ranges into the Trenton. Zones 16 through 20 were included in the "Trenton" of previous reports. Zones 1–6 are probably best developed in southeastern Lee County along the strike from Dickerson Ford and Glass Store to Deep Spring School, while Zones 7–20 are well developed in central Lee County in the vicinities of Harrogate, Tennessee, Hagan, Virginia, and Turkey Cove, Virginia.

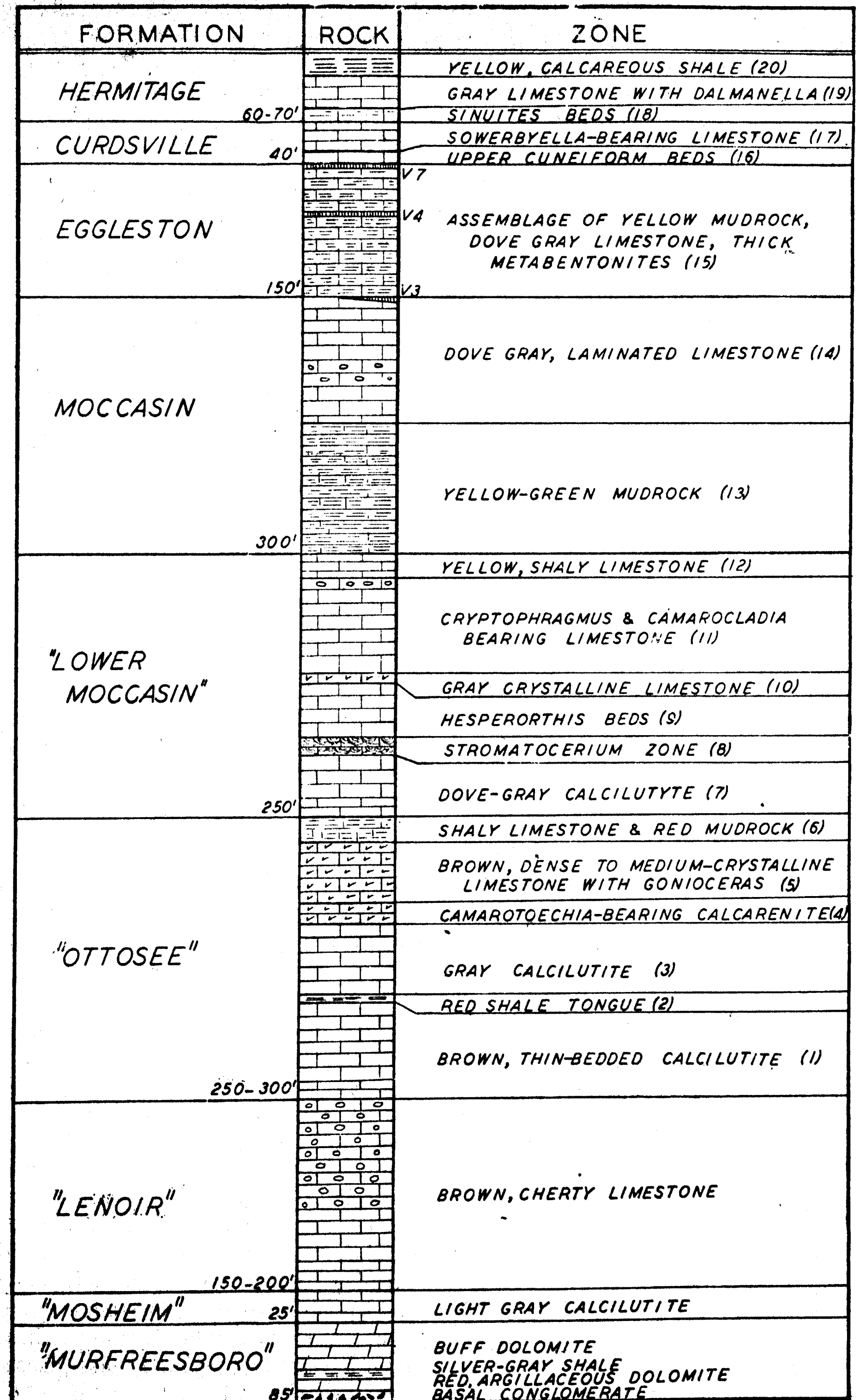


Fig. 3.—Middle Ordovician columnar section for Lee County, Virginia

"OTTOSEE" FORMATION (ZONES 1–6)

The "Ottosee" of Lee County comprises a maximum of 325 feet of brown calcilutite, red shale, gray calcilutite, coarse calcarenite, medium-crystalline limestone, and yellow, shaly, nodular-weathering limestones which constitute Zones 1–6 of the generalized section (Fig. 3 and Table 2). It lies with apparent conformity on the "Lenoir"; a boundary between the formations is drawn arbitrarily where the massive, brown, cherty "Lenoir" gives place to thin-bedded, platy calcilutite. It is overlain conformably by a sequence of beds here referred to as "Lower Moccasin." The best exposures of "Ottosee" are in southeastern Lee County, where the zones are fully developed. Northwestward many of the zones pass into calcilutite and their identity is lost. In central Lee County the lower beds are somewhat argillaceous with ripple marks and fucoid markings. About 150 feet above the base are lenses of coarsely crystalline limestone of the "Holston" lithology; these are believed to represent the most north-

TABLE 2

SUBDIVISIONS OF THE "OTTOSEE" FORMATION

"LOWER MOCCASIN" FORMATION

"OTTOSEE" FORMATION

Zone

6. *Shaly limestone and red mud rock:* In southeastern Lee County, notably Glass Store, and Dickerson Ford, the brown, crystalline *Gonioceras* beds pass upward into shaly, nodular-weathering limestones. At Glass Store this zone is abundantly fossiliferous, with numerous *Dinorthis*, *Sowerbyella*, *Öpikina*, *Multicostella*, and *Strophomena*. The beds pass southeastward into yellow shale and thin beds of limestone with *Öpikina*, *Dinorthis*, *Chasmatopora*, and other genera common to the upper shaly "Ottosee" of Rye Cove. Northwestward these beds pass into a yellow, shaly limestone which becomes red mud rock in central Lee County. Thickness in Lee County about 25 feet.
5. *Brown, dense to medium-crystalline limestone with Gonioceras:* In southeastern Lee County, the granular, *Camarotoechia* beds pass upward into brown, medium-crystalline limestone with a maximum thickness of 75 feet and an average of 50 feet. These beds pass northwestward into dense limestone, locally shaly near the base. In central Lee County, the upper part of this zone forms a massive ledge of dove-gray, calcilutite which becomes *Gonioceras*-bearing near Hagan, Virginia. In southeastern Lee County, *Gonioceras*, etched outlines of *Lophospira*, *Girvanella*, and trilobite fragments are common.
4. *Camarotoechia-bearing calcarenite:* Coarse, friable, somewhat pinkish calcarenite which resembles the "Holston" lithology of Butts. Well exposed at Dickerson Ford, Glass Store, and Deep Spring School, the zone thins northwestward to form discontinuous lenses, the maximum thickness being 35 feet. Characterized by numerous, large *Camarotoechia*.

TABLE 2—*Continued*

Zone

3. *Gray calcilutite:* A maximum of 72 feet of gray, sub-lithographic, massive-bedded limestone with numerous calcite stringers. There are numerous outlines of *Lophospira* near the top. The strata are best exposed at Dickerson Ford and Glass Store, becoming thinner-bedded northwestward; in central Lee County they resemble the underlying Zone 1.
2. *Red shale tongue:* The brown, thin-bedded calcilutite of Zone 1 passes upward into yellow, shaly limestone with a thin tongue of red, calcareous shale. The red shale tongue, well exposed at Dickerson Ford, can be traced along the strike from Yellow Branch Creek to the vicinity of Deep Spring School. It is not exposed in the Glass Store section, and its identity is lost northwestward in central Lee County. Thickness never exceeds a few feet.
1. *Brown, thin-bedded calcilutite:* Brown to dove-gray, dense, thin-bedded limestone lying conformably on the massive, cherty "Lenoir" of Butts. Locally it contains yellow, earthy limestones which weather with a cobbly surface, near the base. Frequently there are thin bands of black chert along bedding planes. It becomes shaly and fucoid-bearing in central Lee County. Maximum thickness is about 140 feet at Dickerson Ford.

"LENOIR" FORMATION

ward spread of the coarse calcarenite (Zone 4) of the southeastern sections. The upper part of the "Ottosee" of central Lee County has 25 feet of yellow, shaly limestone which passes upward into red mud rock. This is persistent, can be mapped, and appears to be at the same horizon as the shaly, nodular-weathering, fossiliferous beds (Zone 6) of the southeastern sections; in both cases the *Stromatocerium* zone of the "Lower Moccasin" lies 65–70 feet above.

Paleontology.—The "Ottosee" of Lee County has few fossils. The most prolific zone is the *Dinorthis* bed in Zone 6 of the Glass Store section. These beds can be traced along the strike to Stickleyville, where they pass into yellow shale with incorporated nodules of limestone, a lithology like that of the upper "Ottosee" of Rye Cove. The fauna collected from the *Dinorthis* bed is listed in Table 3.

The "Ottosee" at Rye Cove consists of two major facies, a lower, brown, crystalline "Holston" lithology and an upper yellow shale and nodular limestone facies. The lower "Ottosee" has *Girvanella*, *Gonioceras*, *Maclurites*, *Carabocrinus* plates, and a prominent zone of associated *Nidulites* and *Sowerbyella aequistriatus*. The lower "Ottosee" of the Glass Store section and Dickerson Ford has abundant *Camarotoechia*; *Tetradium*, *Gonioceras*, *Lophospira*, and *Girvanella*. The upper shaly "Ottosee" of Rye Cove has a large fauna in which the forms listed in Table 4 are abundant.

Fossils are not abundant in the "Ottosee" of central Lee County, where many

TABLE 3

FAUNA FROM THE DINORTHIS BED, ZONE 6

	Locality	
	1	2
Dystactospongia sp.	....	×
Mesotrypa sp.	×	×
Pachydictya sp.	×	×
Camarella sp.	....	×
Dinorthis sp.	....	×
Dinorthis transversa Willard	×	×
Dinorthis sp. *cf. D. quadriplicata* Willard	×	....
Dinorthis sp. *cf. D. atavoides* Willard	×	....
Doloroides sp.	....	×
Fascifera sp.	....	×
Glyptorthis sp. *cf. G. bellarugosa* (Conrad)	....	×
Öpikina sp. *cf. Ö. champlainensis* (Raymond)	....	×
Öpikina magna (Butts)	×	×
Öpikina minnesotensis (Winchell)	×	×
Öpikina sp.	....	×
Multicostella magna Schuchert & Cooper	×	....
Strophomena amploides Butts	×	×
Strophomena medialis Butts	×	×
Strophomena sp.	×	×
Sowerbyella aequistriatus Willard	×	×
Orthoceras sp.	....	×

1. West of road, south of Glass Store, Va.
2. Along road 2½ miles west of Stickleyville, Va.

of the lithologic zones of southeastern sections have passed into dove-gray calcilutites. A small faunule collected along the road at Walnut Hill School west of Gibson Station Post Office is listed in Table 5.

Dinorthis beds (Zone 6) of the Glass Store section, which seem to correspond to the yellow shale and red mud rock of central Lee County can be traced southeastward into yellow shales and limestones which are faunally and lithologically like the upper shaly "Ottosee" of Rye Cove; and (3) the brown, crystalline limestone in the lower "Ottosee" of Lee County sections resembles that in the lower "Ottosee" of Rye Cove.

TABLE 4

FAUNULE FROM UPPER SHALY "OTTOSEE" OF RYE COVE

Receptaculites sp.

Batostoma sevieri Bassler
"*Chasmatopora*" (*Subretopora*) sp.
Cornotrypa inflata (Hall)
Escharopora sp.
Mesotrypa sp.
Pachydictya sp.

Campylorthis deflecta (?) Ulrich & Cooper
Dinorthis transversa Willard
Glyptorthis sp. *cf. G. bellarugosa* (Conrad)
Hesperorthis tricenaria (Conrad)
Lingula sp.
Mimella superba Butts
Multicostella sp. *cf. M. platys* Billings
Leptaena sp. *cf. L. palustris* Willard
Öpikina magna (Butts)
Schizambon cuneatus Willard
Sowerbyella aequistriatus Willard
Strophomena amploides Butts

Bucania sp.

Gonioceras sp.

Bumastus sp.
Illaenus sp.
Illaenus sp. *cf. I. fieldi* Raymond

TABLE 5

FAUNULE FROM "OTTOSEE," WALNUT HILL SCHOOL

Tetradium cellulosum (Hall)

Batostoma sevieri Bassler
Escharopora ramosa Ulrich

Camarotoechia plena (Hall)
Camarotoechia sp.
Ancistrorhyncha sp.
Hesperorthis tricenaria (Conrad)
Glyptorthis sp. *cf. G. bellarugosa* (Conrad)
Öpikina sp.
Strophomena sp.
Zygospira recurvirostris (Hall)

Helicotoma sp.
Lophospira sp.

Gonioceras sp.

Leperditia fabulites (Conrad)
Calliops sp. *cf. C. callicephalus* (Hall)

Correlation.—The lower 250–300 feet of the Lee County "Lowville of Butts" is believed equivalent to the "Ottosee" of Rye Cove. This conclusion is based on the following facts: (1) the "Ottosee" of Lee County is stratigraphically similar to the "Ottosee" of Rye Cove; both units overlie the cherty "Lenoir" and are overlain by dove-gray limestones with *Stromatocerium* and *Cryptophragmus;* (2) the cally like the upper shaly "Ottosee" of Rye Cove; and (3) the brown, crystalline limestone in the lower "Ottosee" of Lee County sections resembles that in the lower "Ottosee" of Rye Cove.

"LOWER MOCCASIN" FORMATION—(ZONES 7–12)

The "Lower Moccasin" includes 250–300 feet of dove-gray limestone (Table 6). A prominent zone of *Stromatocerium*

TABLE 6

SUBDIVISIONS OF THE "LOWER MOCCASIN" FORMATION

"MOCCASIN" FORMATION

"LOWER MOCCASIN" FORMATION

Zone

12. *Yellow-shaly limestone:* This zone includes 25 feet of thin-bedded, shaly limestone with abundant *Rhinidictya nicholsoni*, *Escharopora subrecta*, and *Zygospira recurvirostris*. Well exposed in the vicinity of Harrogate, Tennessee, especially in the railroad cut east of Harrogate.
11. *Cryptophragmus and Camarocladia limestones:* Approximately 100 feet of dove-gray, dense limestone with occasional thin beds of gray, crystalline limestone. Characterized by a zone of *Camarocladia* near the base and *Cryptophragmus antiquatus* throughout. In southeastern Lee County a prominent zone of black chert is at the top.
10. *Gray, crystalline limestone:* Consists of from 3 to 8 feet of coarse, fragmental limestone. Best developed at the base of the quarries southeast of Harrogate, Tennessee.
9. *Hesperorthis beds:* Includes a maximum thickness of 60 feet of dove-gray, dense, platy limestone with a prominent zone of *Hesperorthis tricenaria* and associated *Öpikina* sp. 10 feet above the base. Well developed throughout central Lee County where it invariably forms a prominent marker above the *Stromatocerium rugosum* beds.
8. *Stromatocerium beds: Stromatocerium rugosum* forms a prominent zone throughout Lee County, ranging in thickness from a few inches to a maximum of 20 feet west of Harrogate, Tennessee. This zone is persistent, mappable, and is identifiable in all the Lee County sections and at Rye Cove in Scott County, Virginia.
7. *Dove-gray calcilutite:* Dove-gray, dense, platy limestones. Locally, as at Glass Store, are thin beds of crystalline limestone. Marked at the top by a prominent zone of nodular black chert. Average thickness about 60 feet.

"OTTOSEE" FORMATION

rugosum is 65 feet above the base, and a zone of associated *Hesperorthis tricenaria* and *Öpikina* sp. is 10 feet higher. Beginning 125 feet above the base is a 100-foot zone (minimum) of dense to medium crystalline limestone having *Camarocladia* near the base and *Cryptophragmus antiquatus* throughout. At Harrogate, Tennessee, 25 feet of shaly limestone (Zone 12) with *Rhinidictya* and *Zygospira* overlie the *Cryptophragmus* beds; these shaly strata were not recognized elsewhere. The *Cryptophragmus* zone is about 100 feet in central Lee County; it thickens southeastward. A prominent zone of black chert often forms the top of this zone.

Paleontology.—The "Lower Moccasin" has a number of fossils in addition to those mentioned above. The forms listed in Table 7 were identified.

Correlation.—The strata referred to as "Lower Moccasin" in Lee County can be correlated with the "Lower Moccasin" of Rye Cove on lithological and faunal bases. At Rye Cove, *Stromatocerium* occurs 85 feet above the upper shaly "Ottosee" and *Cryptophragmus* is higher in the section. Use of the term "Lower Moccasin" is undesirable except for correlation purposes; according to Cooper,[24] these beds were not included in the original definition of the Moccasin formation but are actually older than the true Moccasin. New terminology is applied on a later page.

Good exposures of the "Ottosee" and "Lower Moccasin" are in southeastern Lee County, where several sections have been measured. The first section is south of Wallen Creek along the highway about $5\frac{1}{2}$ miles south of Jonesville. This is shown in Table 8.

In summary, the Glass Store section shows several significant features. There is predominance of gray, sublithologic limestone (Zone 3) in the lower part over-

[24] "Moccasin Formation in Southwestern Virginia" (abstr.), *Bull. Geol. Soc. Amer.*, Vol. LIII (1942), p. 1799.

TABLE 7

FAUNULE FROM "LOWER MOCCASIN" OF LEE COUNTY*

Camarocladia sp.
Cryptophragmus antiquatus Raymond

Stromatocerium rugosum Hall

Tetradium cellulosum (Hall)

Batostoma magnapora Ulrich
Batostoma sp.
"*Chasmatopora*" (*Subretopora*) sp. *cf. C. sublaxa* (Ulrich)
Escharopora sp.
Escharopora confluens Ulrich
Escharopora subrecta (Ulrich)
Rhinidictya nicholsoni Ulrich

Ancistrorhyncha sp.
Camarotoechia plena Hall
Fascifera sp.
Hesperorthis tricenaria (Conrad)
Öpikina sp. *cf. Ö. minnesotensis* (Winchell)
Öpikina sp. *cf. Ö. transitionalis* (Okulitch)
Pionodema subaequata (Conrad)
Zygospira recurvirostris (Hall)

Lophospira oweni Ulrich & Scofield
Liospira sp.
Trochonema sp.

Cycloceras sp.
Orthoceras multicameratum Emmons

Illeanus sp.
Trilobite fragments

Leperditia fabulites (Conrad)

* Collections from quarry east of Harrogate, Tenn., and second quarry northeast of Shawnee, Tenn., on Lee County line.

TABLE 8

SECTION AT GLASS STORE, SOUTH OF WALLEN CREEK

Formational Description	Thickness in Feet— Of Unit	To Base of Formation
"Lower Moccasin":		
Cover with dove-gray, platy limestone with zone of black chert at top; *Cryptophragmus* zone	120.0	296.0
Dove-gray, dense, platy limestone	5.0	176.0
Cover	100.0	171.0
Cover with dove-gray, dense, thick-bedded limestone with nodular black chert; *Stromatocerium* in upper beds	43.5	71.0
Gray brown, medium crystalline, thick-bedded limestone with *Gonioceras*	10.5	27.5
Brown, fine-textured, argillaceous limestone	17.0	17.0
"Ottosee":		
Yellow, argillaceous, nodular-weathering limestone with *Dinorthis*, *Öpikina*, *Strophomena*, *Sowerbyella*, others	26.5	301.0
Cover with beds of crystalline limestone	48.0	274.5
Brown, medium-crystalline limestone, weathers to thin, wavy beds	26.5	226.5
Brown, coarsely crystalline, friable limestone with large *Camarotoechia*	19.0	200.0
Gray, thick-bedded calcilutite	72.0	187.0
Cover, immediately south of creek	115.0	115.0
"Lenoir":		
Brown, dense, massive, cherty limestone	....	....

lain by crystalline limestones (Zones 4 and 5), with abundant large *Camarotoechia* near the base. Higher in the section are the fossiliferous, shaly, nodular-weathering *Dinorthis* beds (Zone 6); these are regarded as the top of the "Ottosee." Seventy feet above is a conspicuous black chert bed with a persistent ledge of *Stromatocerium rugosum* at the top (Zones 7 and 8). Higher on the hillside, *Cryptophragmus* (Zone 11) is abundant.

A second section, located on the north side of the Powell River about 5 miles south of Jonesville, is shown in Table 9. This section (Dickerson Ford) compares closely with the Glass Store section. The red shale (Zone 2) roughly 150 feet above the base is not exposed at Glass Store. Beds of the *Dinorthis* horizon (Zone 6) are poorly exposed at Dickerson Ford, where they are represented by beds of shaly limestone. The *Stromatocerium* zone is about 250 feet above the red shale and approximately 70 feet above the shaly, nodular-weathering limestone of Zone 6. A 100-foot zone of *Cryptophragmus* is about 75 feet above the *Stromatocerium* bed.

A third section is exposed along the limb of the Powell Valley anticline south of Deep Spring School on the northwest slope of Wallen Ridge. Inasmuch as cover conceals the contact between the

TABLE 9

SECTION AT DICKERSON FORD, NORTH OF POWELL RIVER

Formational Description	Thickness in Feet— Of Unit	To Base of Formation
"Lower Moccasin":		
Dove-gray, dense limestone with *Cryptophragmus*, weathered outlines of brachiopods and gastropods	106.0	280.3
Cover with ledges of dove-gray, dense limestone	15.9	174.3
Cover with beds of finely crystalline limestone	13.8	158.4
Cover with beds of dove-gray, dense, platy limestone	58.3	144.6
Cover with thin beds of dove-gray limestone carrying *Stromatocerium*	45.0	86.3
Dove-gray, dense, platy limestone	26.5	41.3
Gray-brown, dense to finely crystalline limestone	14.8	14.8
"Ottosee":		
Cover with thin beds of yellow, shaly limestone	19.1	324.0
Brown, medium-crystalline, thin-bedded limestone; forms top of quarry	33.4	304.9
Brown, dense, thick-bedded limestone with *Gonioceras*, outlines of brachiopods and gastropods	13.3	271.5
Cover with ledges of crystalline limestone	19.1	258.2
Brown, crystalline, nodular-weathering limestone with *Girvanella*, *Gonioceras*, trilobite fragments	16.4	239.1
Gray, sublithographic, massive-bedded, limestone with calcite stringers; surface etched and covered with outlines of *Lophospira* sp.	72.1	222.7
Yellow, shaly limestone grading into red, calcareous shale at the top	10.1	150.6
Dove-gray, dense, medium-bedded limestone, weathers thin-bedded; zone of *Tetradium* near base	61.5	140.5
Cover with ledges of thin-bedded limestone	63.6	79.0
Brown, shaly limestone in ledges separated by cover; lower beds weather to irregular, cobbly surface	15.4	15.4
"Lenoir":		
Brown, dense, massive, cherty limestone	Not measured	

"Ottosee" and "Lower Moccasin," it is difficult to divide the section. The brown, crystalline, *Camarotoechia* beds (Zone 4) are well exposed, and a 100-foot zone of *Cryptophragmus* is near the top.

MOCCASIN FORMATION—(ZONES 13 AND 14)

The Moccasin includes two distinct and well-defined facies:

Zone 14. *Dove-gray, laminated limestone:* This zone includes a maximum of 135 feet of dove-gray, thick-bedded laminated limestone with several thin metabentonites near the top and a 25 foot cherty zone about 30 feet above the base. It is best exposed at the Hagan railroad siding, where the entire thickness can be seen. It is partially exposed on Wallen Ridge near Deep Spring School.

Zone 13. *Yellow-green mud rock:* Comprises a maximum of 175 feet of yellow-green calcareous unfossiliferous mud rock. It is generally nonresistant and poorly exposed. The lower beds can be seen near the top of the Wheeler quarry northeast of Gibson Station, the upper beds at Hagan railroad siding, and median ledges on the northwest slope of Wallen Ridge near Deep Spring School.

The Moccasin overlies the "Lower Moccasin" with no evidence of unconformity. It is overlain by the Eggleston formation; a metabentonite zone separates the two formations.

Paleontology and correlation.—Fossils were not collected from this horizon. The beds seem to correspond lithologically to the Moccasin formation of southwestern Virginia.

EGGLESTON FORMATION—(ZONE 15)

The Eggleston is an assemblage of yellow mud rock, dove-gray limestone, and thick metabentonites (Fig. 4). The formation is approximately 150 feet thick in central Lee County; it thickens southeastward to about 175 feet at Wallen Ridge, south of Jonesville. The thickness seems uniform along the strike. A thin metabentonite, exposed at Hagan, Virginia, separates the Eggleston from the underlying Moccasin. A pair of thick metabentonites (3–4 feet) with their associated basal chert zones, separated by 40–50 feet of limestone and yellow, calcite-flecked mud rock, are near the top. The interval between these metabentonites is relatively constant; 40 feet at Hagan, Virginia; 40 feet at Harrogate, Tennessee; 46 feet at Turkey Cove, Virginia; and 49 feet at Wallen Ridge south of Jonesville. The two thick metabentonites seem to correspond to metabentonites *V4* and *V7* of the Rosenkrans classification. Precise correlation of the lower metabentonite has not been made; it seems to correspond to metabentonite *V3* in the lower Eggleston of southwestern Virginia.

The uppermost metabentonite (*V7*) is overlain by 10–15 feet of alternating shale and limestone with intercalated mud rock (Zone 16). The term "cuneiform" has been applied to these beds because of the peculiar right-angle fracture of some of the limestone beds. At Harrogate, Tennessee, and Turkey Cove, Virginia, the thin limestones of the "cuneiform" beds pass upward into a thin, friable calcarenite which forms the base of the crystalline "Trenton" limestones. Inasmuch as these beds are transitional, it is difficult to say whether they should be included in the Eggleston or in the overlying formation. Rosenkrans[25] included the "cuneiform" beds in the "Trenton," drawing his boundary 2 feet below the upper metabentonite, *V7*, where he believed a hiatus occurs. The accordant thickness of beds between the two thick metabentonites in Lee County offers no indication of a hiatus at this horizon. The presence of *Sowerbyella curdsvillensis* and *Dalmanella fertilis* serves to tie these beds with the overlying fossiliferous "Trenton" limestone. In

[25] P. 96 of ftn. 9 (1936).

central Kentucky, the crystalline, *Sowerbyella*-bearing Curdsville limestone lies directly on a thick (3–4 foot) metabentonite which seems to correspond to metabentonite *V7*. It appears best, therefore, to draw the top of the Eggleston at the top of the upper metabentonite and include the "cuneiform" beds in the overlying formation.

Paleontology.—The Eggleston has yielded a number of fossils. The thin, shaly limestones in the lower part carry a fauna with *Rhinidictya* and *Escharopora*. *Tetradium* is common throughout. A limestone above metabentonite *V4* has abundant *Strophomena* and *Öpikina*, while numerous pelecypods and cephalopods are found in the nodular shales and mud rock between the upper metabentonites. The faunules collected from the Eggleston are shown in Table 10.

Correlation.—This is the Eggleston formation or Eggleston facies of authors. The Eggleston was defined by Mathews in 1934 to include "the beds of upper Black River age that are younger than the upper red Moccasin and older than the Trenton." The Eggleston forms a "transition zone" between the Moccasin formation and the limestones of the "Trenton." In areas to the southeast, Rosenkrans[26] has concluded that, since the upper and lower limits of the Eggleston as found in the type locality vary from place to place, the Eggleston is merely a facies of the Moccasin. This paper does not intend to entertain the status of the Eggleston, but, because of its distinctive characteristics in Lee County, it is given equal rank with Moccasin and others. Metabentonites *V7* and *V4* are believed to correspond to the Mud Cave clay and the Pencil Cave clay of the Carters formation in Tennessee.

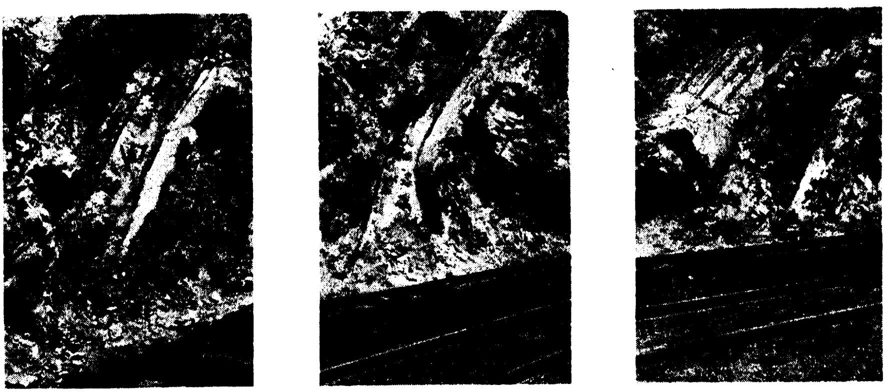


FIG. 4.—Eggleston metabentonites at Hagan, Virginia. *Left:* Upper metabentonite *V7* lying on 4 inches of black, rippled chert. *Center:* Metabentonite *V4:* note changes in thickness introduced by folding. *Right:* Lower metabentonite *V3* (?). Note small fault in center of picture repeating 8 feet of beds.

The Eggleston is nicely exposed at Turkey Cove, Hagan, and on Wallen Ridge south of Jonesville in Virginia, and along the railroad east of Harrogate, Tennessee. Partial sections near Turkey Cove have been described by Bates,[27] and the Wallen Ridge section is given by Rosencrans.[28] The most complete section

[26] *Ibid.*, p. 92.

[27] Pp. 50–51 of ftn. 11 (1939).

[28] Pp. 110–11 of ftn. 9 (1936).

is at Hagan, Virginia, where a continuous section of strata from upper Moccasin to "Catheys-Cannon" can be seen (Table 11).

beds. Very fossiliferous with *Sowerbyella curdsvillensis*, *Dinorthis pectinella*, and *Dalmanella fertilis*. Average thickness 30 feet. Well exposed at Harrogate, Tennessee, and Hagan, Virginia.

TABLE 10

FAUNULES FROM THE EGGLESTON OF LEE COUNTY

	LOCALITIES		
	1	2	3
Stromatocerium sp. (rare)	...	×	...
Lambeophyllum profundum (Conrad)	...	×	...
Tetradium cellulosum (Hall)	×	×	×
Arthroclema sp.	...	...	×
Batostoma sp.	...	...	×
Escharopora sp. *cf. E. confluens* Ulrich	...	×	×
Escharopora subrecta Ulrich	...	×	×
Rhinidictya nicholsoni Ulrich	×	×	×
Doleroides sp. *cf. D. gibbosus* (Billings)	...	...	×
Öpikina sp.	...	×	...
Öpikina sp. *cf. Ö. minnesotensis* (Winchell)	...	×	...
Öpikina sp. *cf. Ö. transitionalis* (Okulitch)	×	...	×
Strophomena sp.	...	...	×
Zygospira recurvirostris (Hall)	×	×	×
Cyrtodonta sp. *cf. C. huronensis* Billings	...	×	×
Helicotoma sp.	...	×	×
Lophospira oweni Ulrich and Scofield	...	×	...
Lophospira sp.	...	×	×
Cameroceras sp.	...	×	...
Cycloceras sp.	...	×	...
Endoceras sp.	...	×	...
Aparchites sp.	...	...	×
Eurychilina subradiata Ulrich	...	...	×
Primitiella sp. (?)	...	...	×
Leperditia fabulites (Conrad)	×	×	×
Calliops sp. *cf. C. callicephalus* (Hall)	...	×	...
Eomonorachus sp.	...	...	×
Illaenus sp.	...	...	×

1. East side of road, Wallen Ridge south of Jonesville, Va.
2. Along road 2½ miles east of Cumberland Gap in Lee County, Va.
3. Along road through Turkey Cove, Va.

CURDSVILLE FORMATION (ZONES 16–17)

The Curdsville, as here designated, includes two zones:

Zone 17. *Sowerbyella-bearing limestone:* Blue-gray, coarsely crystalline limestone in 2-10 inch

Zone 16. *Upper "cuneiform" beds:* A sequence of thin limestones and shale overlying metabentonite *V7*. Average thickness 10–15 feet. Grade upward into a friable calcarenite which forms the base of Zone 17. Well exposed at Turkey Cove and Hagan, Virginia, and Harrogate, Tennessee.

TABLE 11

SECTION ALONG RAILROAD SIDING, HAGAN, VIRGINIA

Formational Description	Thickness in Feet—	
	Of Unit	To Base of Formation
"Catheys-Cannon":		
Blue-gray, medium-crystalline limestone; fossiliferous	230.0	230.0
Hermitage:		
Cover with shale slump	25.0	71.4
Yellow, plastic metabentonite	1.0	46.4
Yellow-brown, platy, fossiliferous shale with *Dalmanella*	10.0	45.4
Blue-gray, medium-crystalline limestone in 2–4 inch beds separated by thin, gray shales; *Sinuites*, *Dalmanella*, *Sowerbyella*	35.0	35.4
Blue, plastic metabentonite	0.4	0.4
Curdsville:		
Blue-gray, coarse-textured limestone in 2–10 inch beds with *Sowerbyella*, *Dinorthis*, *Dalmanella*	30.0	40.2
Gray, dense limestone in thin, nodular layers separated by shale	5.0	10.2
Gray, dense, thick-bedded limestone	3.9	5.2
Yellow, calcareous mud rock	0.8	1.3
Dove-gray, dense limestone in thin beds separated by gray shale	0.5	0.5
Eggleston:		
Yellow-green, micaceous metabentonite	2.0	147.1
Gray, soft, micaceous metabentonite	1.0	145.1
Gray-brown, hard metabentonite (*V7*)	1.0	144.1
Black, rippled chert	0.5	143.1
Dove-gray, dense limestone	6.0	142.6
Yellow, calcareous mud rock	18.0	136.6
Dove-gray, dense limestone	14.0	118.6
Yellow-green metabentonite (*V4*)	3.0	104.6
Black, rippled chert	0.3	101.6
Dove-gray, dense, platy limestone	38.0	101.3
Gray, bentonitic (?) shale	0.5	63.3
Dove-gray, dense, shaly limestone	20.0	62.8
Blue-gray, structureless mud rock	34.0	42.8
Dove-gray, dense, thick-bedded limestone (repeated by small fault)	8.0	8.8
Gray, fine-textured shale	0.3	0.8
Blue-gray, micaceous metabentonite	0.5	0.5
Moccasin:		
Dove-gray, dense limestone	4.0	147.4
Green, shaly metabentonite	0.1	143.4
Dove-gray, dense limestone	20.0	143.3
Blue-gray, shaly metabentonite	0.3	123.3
Dove-gray, dense limestone	30.0	123.0
Dove-gray, dense, laminated limestone with nodular chert	78.0	93.0
Brown-green, massive, mud rock	15.0	15.0
Base of exposure:		

The Curdsville is limited at the base by a 3- to 4-foot metabentonite ($V7$) and is separated from the overlying formation by a varying thickness of blue, hard, shaly metabentonite.

Paleontology.—The Curdsville is highly fossiliferous and is characterized by the association of *Sowerbyella curdsvillensis*, *Dinorthis pectinella*, and *Dalmanella fertilis*. The "cuneiform" beds are sparingly fossiliferous, but slabs with *Sowerbyella* and *Dalmanella* have been collected.

TABLE 12

FAUNULES FROM THE CURDSVILLE OF SOUTHWESTERN VIRGINIA

	LOCALITIES		
	1	2	3
"*Chasmatopora*" (*Subretopora*) sp.	×	....	×
Hallopora sp.	....	....	×
Prasopora sp.	....	×	×
Rhinidictya sp.	....	×	....
Dalmanella fertilis Bassler	×	×	×
Dinorthis pectinella (Emmons)	×	×	×
Hesperorthis tricenaria Conrad	×	×	×
Rafinesquina alternata	×	....	×
Sowerbyella curdsvillensis (Foerste)	×	×	×
Zygospira recurvirostris Hall	....	....	×
Cyrtolites ornatus Conrad	....	....	×

1. Railroad cut southeast of Harrogate, Tenn.
2. Railroad siding east of Hagan, Va.
3. East side of Big A Mountain, ¼ mile north of Fullers Corner, Russell County, Va. (collected by Marshall Kay).

Correlation.—The fauna and lithology of this formation resemble those of the Curdsville formation in areas to the north and west. The Curdsville apparently has not been recognized in the northwest belt of the Appalachian Valley, although fossil lists invariably carry Curdsville fauna. The widespread nature of the Curdsville in southwestern Virginia is indicated by its presence as far north as Big A Mountain near Honaker, Virginia. A faunule collected from the east side of Big A Mountain is compared in Table 12 with that from Lee County, Virginia.

Good exposures of the Curdsville are at Harrogate, Tennessee, Hagan, Virginia, and Wallen Ridge south of Jonesville, Virginia. It was not recognized by earlier workers in Wise County,[29] but its presence as far northeast as Big A Mountain, Honaker, Virginia, indicates that its deposition was not limited to the Lee County area. The section at Hagan, Virginia, has been included in Table 11; a section at Harrogate, Tennessee, is shown in Table 13.

HERMITAGE FORMATION (ZONES 18–20)

The Hermitage consists of the following:

Zone 20. *Yellow calcareous shale:* Includes a maximum of 30 feet of yellow, calcareous shale with thin beds of fossiliferous limestone carrying abundant *Dalmanella*. A 12-inch bed of soft, yellow, metabentonite is 10 feet above the base. These beds are nonresistant and yield readily to slumping.

Zone 19. *Gray limestone with Dalmanella:* Gray, medium-crystalline, thin to medium-bedded limestone with thin shale partings. Marked by the abundance of *Dalmanella fertilis*. Thickness 15–20 feet. Passes upward into a massive ledge of yellow, hard, calcareous shale which forms the base of Zone 20.

Zone 18. *Sinuites Beds:* Comprises 10–15 feet of gray, fine-textured limestone in thin beds separated by thin layers of gray, siliceous shale. Characterized by an abundance of *Sinuites cancellatus* and *Dalmanella fertilis*.

The Hermitage is separated from the underlying Curdsville by a bed of hard, blue metabentonite. Contact with the overlying "Catheys-Cannon" is not clearly shown in the limited exposures examined. The Hermitage is exposed along the road 2½ miles east of Cumberland Gap, Tennessee; along the railroad at Harrogate, Tennessee; and along the sid-

[29] Stose *et al.*, pp. 22–28 of ftn. 10 (1923).

ing at Hagan, Virginia (see Tables 11 and 13).

Paleontology.—The Hermitage is very fossiliferous, with a zone of *Sinuites* near the base and *Dalmanella* common throughout. The fauna collected from the Lee County exposures is shown in Table 14.

Correlation.—Lithologically and faunally these beds correspond to the post-Curdsville portion of the Hermitage of Tennessee. According to C. W. Wilson, Jr.,[30] the lower part of the Hermitage formation in the central basin of Tennessee can be separated from the overlying part of the Hermitage as a distinct lithologic and faunal unit which he considers to be of Curdsville age. This not only serves to strengthen the correlation

TABLE 13

SECTION ALONG RAILROAD EAST OF HARROGATE, TENNESSEE

Formational Description	Thickness in Feet—	
	Of Unit	To Base of Formation
Hermitage:		
Yellow to reddish-brown shale	7.6	48.7
Yellow, crumbly metabentonite	0.6	41.1
Yellow, thin-bedded shale	2.5	40.5
Yellow-drab, calcareous shale in one massive bed, breaks to thin yellow shale with intercalated beds of fossiliferous limestone	6.0	38.0
Blue-gray, medium-crystalline limestone, thin to medium bedded with thin shale partings; fossiliferous with *Dalmanella*	15.0	32.0
Gray, fine-textured, thin-bedded limestone separated by gray, siliceous shale; abundant *Sinuites*	15.0	17.0
Blue, hard, shaly metabentonite, breaks to angular fragments	2.0	2.0
Curdsville:		
Blue-gray, medium-crystalline limestone in medium beds with thin shale partings; with *Dinorthis*, *Sowerbyella*, and *Dalmanella*	10.0	45.6
Blue, fissile shale	0.5	35.6
Blue-gray, crystalline limestone in medium beds, highly fossiliferous with *Dinorthis* and *Sowerbyella*	25.5	35.1
Gray to brown, dense limestone passing upward into a brown, friable calcarenite which forms the base of overlying limestone	3.0	9.6
Dove-gray, dense limestone in nodular beds separated by blue-gray shale	1.2	6.6
Dove-gray, dense limestone in medium beds, weathers shaly	4.2	5.4
Drab-green, hard, angular shale	1.2	1.2
Eggleston:		
Gray-green, crumbly metabentonite with abundant biotite flakes ($V7$)	3.0	54.0
Black chert	0.3	51.0
Brown, dense, platy limestone with *Leperditia* and *Camarotoechia*	3.2	50.7
Yellow-blue, calcareous mud rock	35.0	47.5
Dove-gray, dense, platy limestone	1.3	12.5
Yellow, micaceous metabentonite ($V4$)	3.0	11.2
Black chert	0.2	8.2
Dove-gray, dense limestone	8.0	8.0
Base of exposure:		

[30] "Curdsville Limestone Zone of Hermitage in Central Tennessee" (abstr.), *Bull. Geol. Soc. Amer.*, Vol. XLIX (1938), p. 1923.

of the Virginia and Tennessee sections but emphasizes the widespread nature of the Curdsville formation in the Appalachian Valley.

OVERLYING ROCKS

The "Catheys-Cannon" consists of blue-gray, granular, even-bedded, fossiliferous limestone with a maximum thickness of about 500 feet. No detailed study was made, but casual examination indicated that any division in Lee County should be made on faunal rather than on lithological criteria. Two formations, the Catheys and Cannon, have been recognized in Wise County,[31] and sections have been described in the Wise County report. Good exposures are rare in Lee County, but a continuous sequence of 230 feet is well shown at Hagan, Virginia.

TABLE 14

FAUNULES FROM THE HERMITAGE OF LEE COUNTY

	Localities		
	1	2	3
Hallopora sp.	×	....	×
Prasopora sp.	×	....	×
Prasopora falesi Hall	×	....	×
Dalmanella fertilis Bassler	×	×	×
Hesperorthis tricenaria (Conrad)	....	×	....
Heterorthis clytie Hall	....	×	....
Rafinesquina hermitagensis Bassler	×	×	×
Rhynchotrema increbescens (Hall)	×	×	×
Sinuites cancellatus Hall	×	....	×
Modiolodon sp.	×	×	....
Calliops sp.	....	×	....
Crinoid stems	....	×	....

1. Railroad cut southeast of Harrogate, Tenn.
2. Along road 2½ miles east of Cumberland Gap, Tenn.
3. Railroad siding, Hagan, Va.

Application of the Cooper and Prouty Terminology to the Lee County Section

Much of the terminology proposed by Cooper and Prouty[32] for the Middle Ordovician of Tazewell County, Virginia, can be applied to the Lee County section. Several of the zones of the Tazewell County classification have not been recognized in Lee County; they may be missing, or their identity is lost as they pass into calcilutites.

The "Murfreesboro" of Lee County corresponds to the Blackford member of the Cliffield formation which includes Zones 1, 2, and 3 of the Tazewell section. The Blackford lies unconformably on the Beekmantown dolomite; a prominent bed of basal clastics is present. The "Mosheim" of Lee County is the first calcilutite (Zone 4), or the Five Oaks member of the Cliffield formation, and the "Lenoir" is probably, at least in part, equivalent to the Lincolnshire member. The brown calcilutite (Zone 1 of the "Ottosee" of Lee County) is stratigraphically similar to the Ward Cove limestone, while the gray calcilutite (Zone 3 of Lee County) with the numerous *Lophospira* seems to correspond to the *Lophospira* beds and the second calcilutite of the Peery limestone. The *Camarotoechia*-bearing calcarenite and the overlying *Gonioceras* beds probably correspond to the second coarse-grained limestone of the Tazewell section (lower Benbolt), while the shaly, nodular *Dinorthis* beds of the Glass Store section and the upper shaly part of the "Ottosee" of Rye Cove are considered equivalent to the upper Benbolt. The Gratton limestone of Tazewell County is then represented by a dove-gray calcilutite (Zone 7) in Lee County. Zones 8, 9, and 10 of Lee County can be correlated with the

[31] Stose *et al.*, pp. 22–28 of ftn. 10 (1923).

[32] P. 884, Fig. 3 of ftn. 3 (1943).

Wardell formation of Tazewell County, which includes the *Stromatocerium rugosum* beds at its base. Strata equivalent to the Bowen formation were not recognized in Lee County. The *Cryptophragmus* and *Camarocladia* beds which form the upper part of the "Lower Moccasin" of Lee County are correlated with the Witten limestone of Tazewell County. Two series have been recognized: the Highbridge series, including the Camp Nelson, Oregon, and Tyrone formations, and the Lexington series, including the Curdsville, Logana, Jessamine, Benson, Brannon, Woodburn, Perryville, and Cynthiana in ascending order. This paper is concerned primarily with five formations: Camp Nelson, Oregon, Tyrone, Curdsville, and Logana ("Hermitage"). All these formations have their type localities in the central Kentucky area (see Fig. 5). The name "Hermitage" has been borrowed from Tennessee by some to replace the earlier name "Logana."

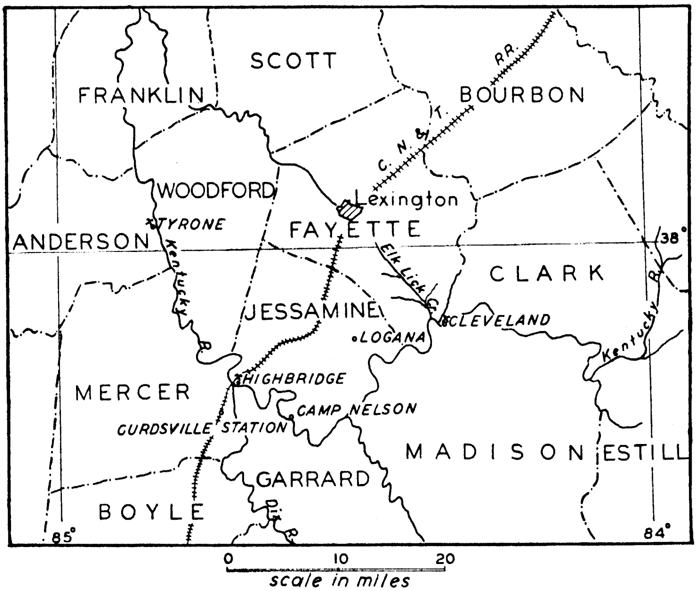


FIG. 5.—Map showing location of sections in central Kentucky

The Moccasin and Eggleston formations of Lee County are believed to be the same as the Moccasin and Eggleston of Tazewell County, while the overlying Curdsville and Hermitage have been included in the Martinsburg shale of northern Virginia.

STRATIGRAPHY OF THE CENTRAL KENTUCKY SECTION

Introduction

Middle Ordovician formations are exposed along the bluffs of the Kentucky River south of Lexington, Kentucky.

Stratigraphic Sequence

"Highbridge Series"

Camp Nelson Formation

The Camp Nelson was named for exposures along the Kentucky River, especially in the vicinity of Camp Nelson,

where approximately 315 feet is exposed above water level. The rock is largely of dense, dove-gray limestone with intercalated streaks of dolomitic material which give a mottled appearance on fresh surface and result in a honeycombed condition upon weathering. The lower 100 feet of exposed limestone is massively bedded and sparingly fossiliferous. The succeeding ledges are thinner bedded and locally contain shaly beds which weather to form deep recesses. The formation becomes increasingly shaly near the top, and locally, as at Highbridge, a 3-foot bed of blue, calcareous mud rock is at or near the top. The base of the Camp Nelson is not exposed, but well sections indicate a thickness of as much as 480 feet. It is overlain by the Oregon formation; the relation of these two formations will be discussed later.

TABLE 15

FAUNULES FROM THE CAMP NELSON FORMATION

	Locality		
	1	2	3
Tetradium cellulosum (Hall)	....	×	....
Escharopora sp. *cf. E. confluens* Ulrich	....	×	....
Mesotrypa sp.	×	×	×
Unidentified Bryozoa	×	×	×
Ancistrorhyncha sp.	×	×	×
Camarotoechia plena (Hall)	×	....	×
Doleroides sp.	×	....	×
Glyptorthis sp.	×	....	×
Multicostella(?) sp. *aff. M. platys* (Billings)	×	....	....
Öpikina sp. *cf. Ö. minnesotensis* (Winchell)	×	×	×
Öpikina sp. *cf. Ö. transitionalis* (Okulitch)	×	....	....
Strophomena sp.	×	....	....
Cycloceras sp.	....	....	×
Gonioceras sp.	×	×	×
Helicotoma sp.	....	×	....
Liospira sp.	×	....	....
Lophospira sp.	×	....	×
Cryptophragmus antiquatus Raymond	×	×	×
Camarocladia sp.	×	×	×
Tentaculites sp.	×	....	×
Calliops sp.	×	....	....
Isotelus sp. (cephalon)	×	....	....
Leperditia fabulites (Conrad)	×	×	×
Cystid plates	....	×	....

1. Along north roadside above Camp Nelson bridge.
2. Shaly beds along north river bluff, Highbridge Park.
3. Along road north of Brooklyn Bridge.

Paleontology.—Fossils are not abundant in the Camp Nelson. The upper part contains scattered *Gonioceras* and *Maclurites*, while the shaly beds have a few species of brachiopods and other fossils. *Camarocladia* is common and *Cryptophragmus antiquatus* forms a prominent zone (65 feet or more) in the upper 90 feet. *Stromatocerium* occurs only sparing-

ly. The faunule collected by the writer is listed in Table 15.

OREGON FORMATION

The Oregon was named for exposures at Lock Number 6, near Oregon, in southern Woodford County, Kentucky, where it consists of 35 feet of heavy, even-bedded, highly magnesian, cream-colored dolomite. The rock is unfossiliferous and breaks with a conchoidal fracture. The Oregon overlies the Camp Nelson. A disconformity is suggested by local variations in the base of the Oregon. At Elk Lick Falls two heavy beds of cream, dolomitic limestone, prominently jointed, lie with sharp contact on the underlying Camp Nelson; at Highbridge a 3-foot bed of blue mud rock is near their contact; at other sections the limestones of the Camp Nelson type seem to grade into those of the Oregon type. The Oregon is overlain conformably by the Tyrone. A sequence of 8–10 feet of beds of the mixed Oregon and Tyrone types lies below the lowest metabentonite of the Tyrone.

TYRONE FORMATION

This is the "Birdseye" limestone of Linney. It was named for exposures near Tyrone, Anderson County, Kentucky, where about 75 feet is exposed. The Tyrone is largely of dove-gray limestone which breaks with a conchoidal fracture and exhibits on fresh surface, calcite facets which have suggested the name "Birdseye." The Tyrone has three metabentonites, two of which are very prominent. Associated with the two upper metabentonites are 3- to 4-inch beds of black chert formed at the contact with the subjacent limestone. The lowest metabentonite is at or near the base of the Tyrone. The second metabentonite is 55 feet above, while the third separates the Tyrone from the overlying Curdsville. The lower metabentonite varies in thickness from a thin clay seam to 18 inches; the second is typically from 30 inches to 3 feet; the upper varies from 3–4 feet. The metabentonites are white to green in color. The lower 55 feet of limestone is dove-gray, sublithographic, and weathers with a whitish surface.

TABLE 16

FAUNULES FROM THE TYRONE FORMATION

	Locality	
	1	2
Escharopora sp. *cf. E. confluens* Ulrich	×	×
Rhinidictya nicholsoni Ulrich	×	×
Batostoma sp.	×	×
Dermatostoma tyronensis Foerste	×	×
Tetradium cellulosum (Hall)	×	×
Tetradium sp. *cf. T. racemosum* Raymond	×	×
Strophomena sp. *aff. S. dignata* Fenton	×	×?
Strophomena sp. *cf. S. plattinensis* Fenton	....	×
Actinoceras sp.	....	×
Endoceras sp.	....	×
Helicotoma planula Salter	×	×
Hormotoma gracilis (Hall)	×	×
Lophospira oweni Ulrich and Scofield	×	×
Raphistoma denticulata Ulrich	....	×
Sactoceras tyronensis Foerste	×	×
Bathyurus extans Hall	....	×
Bathyurus spiniger Hall	×	....
Drepanella crassinoda Ulrich	....	×
Isochilina sp.	....	×

1. Along Elk Lick Creek west of highway south of Lexington, Ky.
2. Kentucky Stone Quarry, Highbridge, Ky.

Conspicuous ledges carry a small fauna with numerous *Tetradium*. The 20 feet of limestone separating the two thick metabentonites always includes a few feet of yellow-green, calcite-flecked mud rock. The upper limestone beds have a small amount of nodular gray chert.

Paleontology.—The Tyrone is not very fossiliferous, although there are several conspicuous fossil horizons. The shaly layers in the lower 55 feet have a few

brachiopods and associated fossils; the black chert at the base of the second metabentonite is filled with silicified outlines of gastropods and cephalopods; the limestone above the second metabentonite carries numerous fossils; the mud rock in the upper 20 feet has *Rhinidictya* and *Zygospira*. The faunules collected from the Tyrone are shown in Table 16.

"LEXINGTON SERIES"

CURDSVILLE FORMATION

The Curdsville, named for exposures near Curdsville Station, Mercer County,

FIG. 6.—Typical cross-bedding in the basal Curdsville as shown in sections along the railroad south of Highbridge, Kentucky.

Kentucky, comprises 20 feet of gray, coarsely crystalline limestone with some nodular chert. The Curdsville lies with apparent conformity on the upper metabentonite of the Tyrone. The basal Curdsville is "rolled" and slightly chertified at the contact with the metabentonite. Two feet above the base, the crystalline limestones locally show cross-bedding, indicative of shallow-water deposition (Fig. 6). The upper beds typically weather to a red residual clay filled with nodules of white chert. A thin metabentonite has been recognized near the top of the Curdsville in some localities.[33]

[33] D. M. Young, "Bentonitic Clay Horizons and Associated Chert," *Univ. Ky. Research Club, Bull. 6* (1940), pp. 27–31.

The presence of a disconformity at the base of the Curdsville is not confirmed by present studies. The Curdsville, near the type locality, lies in direct contact with the thick metabentonite at the top of the Tyrone (Fig. 7). This metabentonite is persistent over a large area and can be

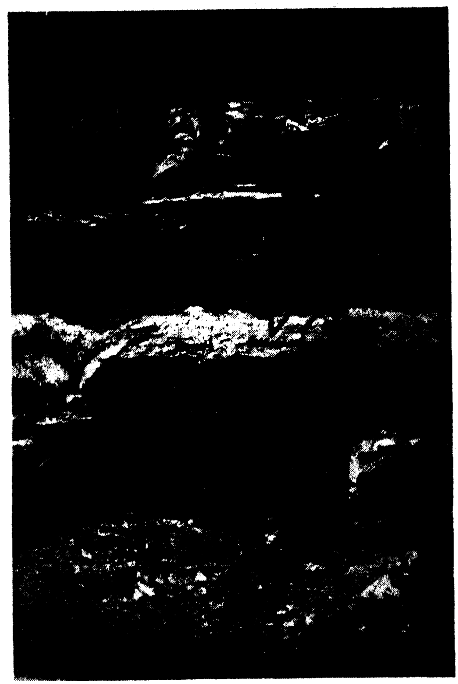


FIG. 7.—Curdsville limestone overlying the upper metabentonite in the Tyrone. Exposure 1½ miles south of Highbridge, Kentucky, along the railroad.

traced for considerable distance with little diminution in thickness. It seems impossible for an unconformity of any magnitude to exist at a metabentonite horizon without removal of the clay. Local reported absence of the clay with "welded" contact between the Curdsville and Tyrone can be attributed to agitation by wave action at or near wave base.

Paleontology.—The Curdsville is characterized by the association of *Sowerbyel-*

la curdsvillensis and *Dinorthis pectinella*. Fossils collected at the type locality are listed in Table 17.

LOGANA ("HERMITAGE") FORMATION

The name "Logana" was given by A. M. Miller[34] to 10 feet of argillaceous limestone and shale overlying the Curdsville and underlying the Hermitage. In 1906[35] the definition was expanded to include the beds between the Curdsville and Wilmore (Jessamine). Later, Foerste and Miller[36] stated that the Logana is the same as the Hermitage of Tennessee, and the name "Hermitage" was applied to 35 feet of beds of which the original Logana formed the basal 10 feet. Raymond[37] continued to use the name "Logana" for the 25–35 feet of clayey limestone and shale immediately above the Curdsville. Inasmuch as the Hermitage has its type locality in Tennessee and since, according to Wilson,[38] the Hermitage of central Tennessee may contain strata of Curdsville age, retention of the name "Logana" for strata immediately above the Curdsville of Kentucky is urged.

The Logana ("Hermitage") of Kentucky consists of 35 feet of gray limestone in thin beds separated by siliceous gray shales. The formation is abundantly fossiliferous and is marked by the association of *Dalmanella fertilis*, *Heterorthis clytie*, *Prasopora falesi*, *Rafinesquina hermitagensis*, and *Lophospira* sp. The lower 10 feet is more siliceous and thinner-bedded than the succeeding 25 feet. The Logana conformably overlies the Curdsville and underlies the Jessámine.

TABLE 17

FAUNULE FROM THE CURDSVILLE FORMATION

Columnaria halli Nicholson
Lambeophyllum profundum (Conrad)

Dinorthis pectinella (Emmons)
Hesperorthis tricenaria Conrad
Platystrophia extensa McEwan
Rafinesquina sp.
Rhynchotrema increbescens (Hall)
Sowerbyella curdsvillensis (Foerste)

Cyrtodonta sp.
Cyrtolites ornatus Conrad
Vanuxemi sp.

Bathyurus spiniger Hall

Crinoid stems

OVERLYING ROCKS

The Logana is overlain by the sequence Jessamine, Benson, Brannon, Woodburn, Perryville, and Cynthiana. These are of middle and upper Trenton age. No study was made of these units, and correlations are, of necessity, from other sources.

CORRELATION WITH THE LEE COUNTY SECTION

The *Cryptophragmus* beds of the upper Camp Nelson can be correlated with the *Cryptophragmus* beds of the "Lower Moccasin" of Lee County or the Witten formation of Tazewell County, Virginia. The Tyrone of Kentucky is like the Eggleston of Virginia; the two thick metabentonites near the top of the Tyrone seem to correspond to a pair near the top of the Eggleston. The two formations

[34] *Ky. Geol. Surv. Bull. 2* (1905), pp. 9, 19.

[35] A. F. Foerste, *Ky. Geol. Surv. Bull. 7* (1906); Miller, *Ohio Nat.*, Vol., VI (1906), pp. 447–48.

[36] Foerste and Miller, *Ky. Geol. Surv.*, Vol. I, Part I (4th ser., 1913).

[37] P. E. Raymond, "Trenton of Central Tennessee and Kentucky," *Bull. Geol. Soc. Amer.*, Vol. XXIII (1922), p. 573.

[38] P. 1923 of ftn. 29 (1938).

have a similar fossil assemblage, and they are lithologically similar—the limestones in the Eggleston are like those in the Tyrone, and the beds of calcite-flecked mud rock in the Tyrone resemble those in the Eggleston. Correlation of the Oregon is uncertain, since it is unfossiliferous. It seems conformable with the overlying Tyrone. The upper Moccasin of Virginia is conformable with the Eggleston. Hence, the Oregon and Moccasin are in part equivalent; they occupy the same stratigraphic position.

The Curdsville of Kentucky is lithologically, faunally, and stratigraphically like the strata referred to as Curdsville in Lee County. The "cuneiform" beds at the base of the crystalline limestone facies in Lee County are believed to be a deeper-water facies of the lower Curdsville.

The Logana ("Hermitage") of Kentucky is equivalent to the Hermitage of the Lee County section and can be correlated with the post-Curdsville portion of the Hermitage of Tennessee.

Correlation of the overlying rocks has not been attempted by the writer. The correlations that appear below are from previous literature. The Jessamine, which is the zone of *Prasopora simulatrix*, has not been recognized in central Tennessee. Inasmuch as the Jessamine is not sharply separated from either the underlying Logana or the overlying Benson in Kentucky, its identity in the Appalachian region may be obscured. The Benson has been correlated with the Bigby of Tennessee; in fact, the Bigby formation of Ulrich (Columbia folio) is regarded as including the Benson, Brannon, and Woodburn.[39] The Bigby has not been recognized in Virginia, but it has been suggested by R. E. Bassler[40] that the Bigby may pass laterally into the Cannon of Virginia and Tennessee. According to Wilson,[41] the Bigby and Cannon limestones are of contemporary age, the Bigby-Cannon unit being composed of several facies, with the Bigby facies thickest along the western part of the central basin and the Cannon facies thickest along the eastern part. The overlying Catheys would then seem to correrespond to the Perryville and Cynthiana.

Convergence from Virginia to Kentucky

An interesting convergence pattern can be worked from Virginia to Kentucky, using the preceding correlations. The minimum thickness of the *Cryptophragmus* beds in Lee County (100 feet) thins to a minimum of 65 feet in the upper Camp Nelson. The Eggleston thins from 150 feet at Hagan, Virginia, to 75 feet in the Tyrone of Kentucky. The interval between the corresponding metabentonites (*V4* and *V7*) thins from 40 feet at Harrogate, Tennessee, to 20 feet in Kentucky. The Curdsville thins from 40 feet to 20 and the Hermitage-Logana from 70 to 35.

The convergence of the *Cryptophragmus* beds, the Eggleston-Tyrone, Curdsville, and Hermitage-Logana from Lee County to central Kentucky is similar. The interval between the *Cryptophragmus* beds and the Eggleston-Tyrone decreases much more rapidly, suggesting the presence of an unconformity within the interval represented by the Oregon. The underlying *Cryptophragmus* beds as now known, converge about 35 per cent;

[39] A. F. McFarlan, "Paleontology of Kentucky," *Ky. Geol. Surv.*, Vol. XXXVI (6th ser., 1931), p. 50.

[40] "Stratigraphy of the Central Basin of Tennessee," *Tenn. Geol. Surv. Bull. 38* (1932), pp. 86–87.

[41] "The Bigby, Cannon, and Cathey Formations in Central Tennessee" (abstr.), *Tenn. Acad. Sci. Jour.*, Vol. XVI, No. 2 (1941), p. 256.

the Eggleston-Tyrone convergence is about 50 per cent. An anticipated convergence for the Oregon should be about 40 per cent. Assuming that no break exists at the base of the Moccasin in Virginia and granting that the Moccasin and Oregon are stratigraphically equivalent, the 300 feet of Moccasin should thin northward to about 180 feet. The Oregon has a maximum thickness of 35 feet, representing a thinning of about 90 per cent. This abrupt change in convergence suggests a hiatus at the base of the Oregon. If a hiatus exists at the base of the Moccasin in Virginia, then the break in the Kentucky section is considerably greater than indicated by the convergence (see Fig. 8).

More accurate dating of the St. Peter sandstone is made possible through convergence studies. Freeman[42] has recognized St. Peter sand in deep wells of Kentucky between the Knox dolomite (Beekmantown) and the Camp Nelson limestone. Isopachal studies by the same author, on the interval between the Pencil Cave clay (*V4*) and the top of the Knox dolomite, show northwest thinning across southern and central Kentucky. This thinning is supported by present convergence studies. The St. Peter is definitely younger than the Knox and older than the Camp Nelson. If the *Cryptophragmus* beds in the upper Camp Nelson are Pamelia in age, then the lower Camp Nelson is pre-Pamelia or Chazyan. The St. Peter sandstone lies above Beekmantown beds and below pre-Pamelia (Chazyan), indicating middle or lower Chazyan age.

Northwestward thinning of middle Ordovician units from southwestern Virginia to the Jessamine dome does not imply the existence of a Cincinnati axis during this time, inasmuch as certain Trenton structures were transverse to that line of reference.[43]

CLASSIFICATION OF VIRGINIA AND KENTUCKY SECTIONS

Correlation of the Virginia and Kentucky sections with the standard section of New York can be accomplished through Pennsylvania. According to Marshall Kay,[44] the Loysburg of central Pennsylvania, classed as Chazyan, is similar to the "Mosheim" and "Murfreesboro" of West Virginia and Virginia. The Hatter has the stratigraphic position of the "Lenoir" and "Ottosee," the upper Hatter (Hostler member) containing forms common to the upper "Rye Cove Ottosee" and the Benbolt of Tazewell County. The *Cryptophragmus*-bearing Benner, classed as Pamelia, forms the lowest "Moccasin" in Virginia, younger Black River (Curtin) being absent. The lower Trenton Nealmont lies with great regional unconformity and local marked disconformity on older beds. The Nealmont passes southward into red shale-bearing Moccasin and overlying Eggleston; metabentonites in the upper part of the basal Oak Hall member of the Nealmont have been correlated with those in the Eggleston. The middle Trenton Salona formation has *Cryptolithus tesselatus* and a zone of *Sinuites cancellatus* near the base. The lower Salona is correlated with the Shoreham of New York, the upper Salona with the Denmark. The Coburn limestone and the Antes black shale of the upper Trenton seem equivalent to

[42] Pp. 1836–43 of ftn. 2 (1939).

[43] McFarlan, "Stratigraphic Relationships of the Lexington, Perryville, and Cynthiana (Trenton) Rocks in Central Kentucky," *Bull. Geol. Soc. Amer.*, Vol. XLIX (1938), pp. 989–96; "Cincinnati Arch and Features of Its Development," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXIII (1939), pp. 1847–52.

[44] "Middle Ordovician of Central Pennsylvania," *Jour. Geol.*, Vol. LII (1944), pp. 1–23, 97–116.

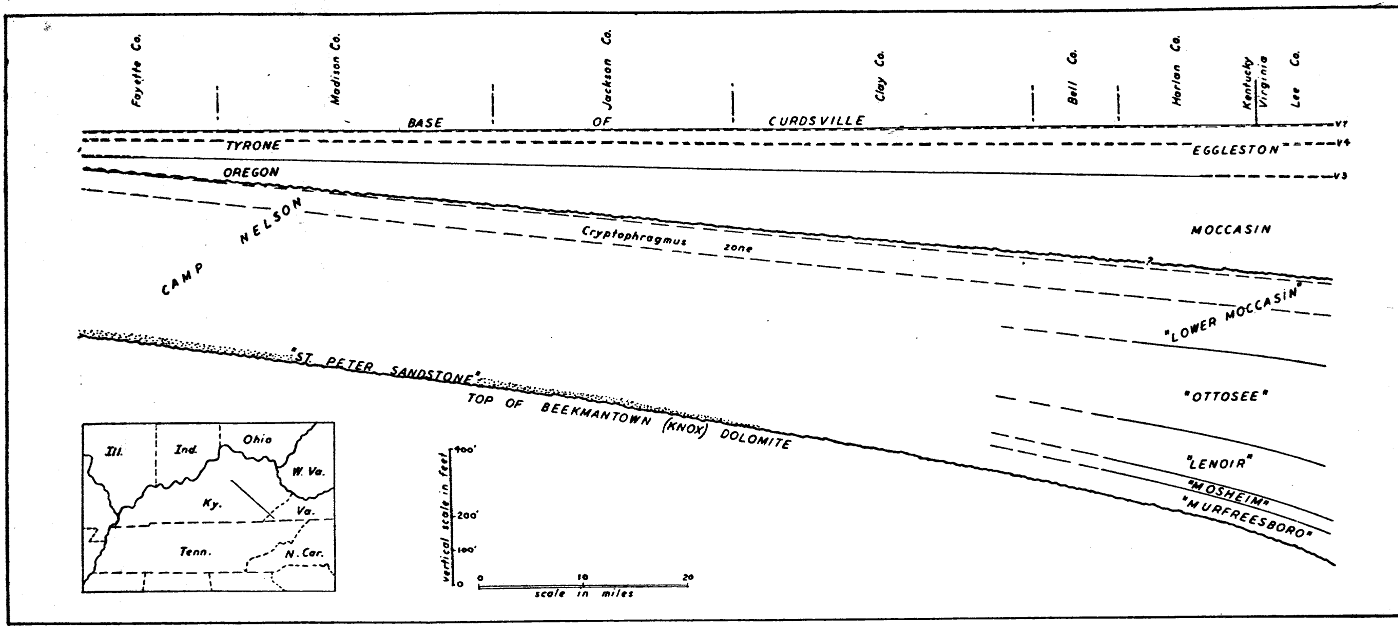


FIG. 8.—Diagram illustrating convergence of middle Ordovician units from Lee County, Virginia, to central Kentucky (data partly from Freeman)

the Cobourg and Holland Patent (Upper Utica) beds in New York.

Kay[45] correlated the Benner limestone, containing the *Cryptophragmus* beds in Pennsylvania, with the Pamelia, and the succeeding Curtin with the Lowville. Inasmuch as the *Cryptophragmus* is present in the Chaumont of New York, the Curtin must be older than the Chaumont. An alternative correlation has since been suggested.[46] Metabentonite *F* is the thickest in the Black River of Pennsylvania and that in the lower Low-

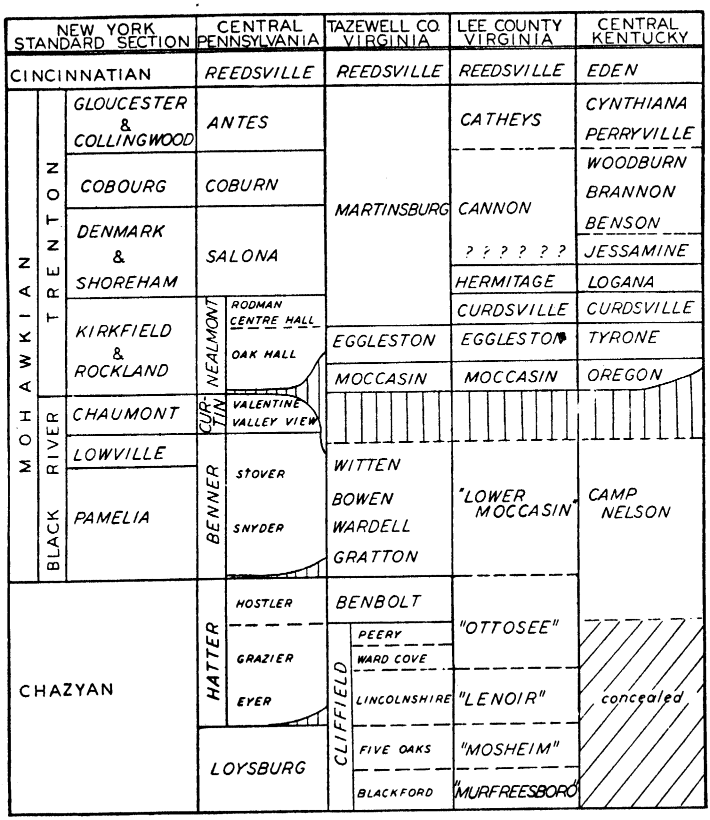


FIG. 9.—Middle Ordovician correlation table

Cryptophragmus ranges from the Pamelia into the Lowville in New York and Ontario. It was thought that metabentonite *A* in the top of the lower Curtin (Valley View) might correspond to that in the lower Lowville of New York and that, inasmuch as metabentonites are absent in the upper Curtin (Valentine) but present in the Chaumont of New York, the Curtin must be older than the Chaumont. ville thickest in New York. If these are correlated, the *Cryptophragmus* zone remains in the corresponding units, the Benner being equivalent to the Pamelia and the Lowville. The Curtin (Valley View) contains several metabentonites, as does the lower Chaumont (Leray). Thus the correlation of the Benner lime-

[45] *Ibid.*, pp. 22–23.

[46] Kay, personal communication (May, 1944).

stone with the Pamelia and the Lowville, and the Curtin with the Chaumont, seems better than the earlier suggestion.

The *Cryptophragmus* beds of the upper Camp Nelson of Kentucky, the "Lower Moccasin" of Rye Cove and Lee County, Virginia, and the Witten limestone of Tazewell County, Virginia, correspond to the upper Benner formation of Pennsylvania, which has been correlated with the Pamelia and Lowville of New York. The lower Trenton Nealmont of Pennsylvania becomes the Moccasin and Eggleston of Virginia and probably includes strata equivalent to the Curdsville of Kentucky near the top. A regional unconformity may be at the base of the Moccasin of Virginia and the Oregon of Kentucky, strata equivalent to the Curtin (Chaumont) being absent. A metabentonite corresponding to metabentonite *F* of the upper Benner has not been recognized in the upper *Cryptophragmus* beds of Virginia. Lower Trenton age of the Eggleston and Tyrone is indicated by their conformity with the Curdsville, which has been correlated with the Kirkfield (Hull) of Ontario and New York on the basis of its echinoderm fauna. *Cryptolithus tesselatus* in the Logana ("Hermitage") of Kentucky correlates that formation with the Salona of Pennsylvania and the Shoreham of New York. The *Sinuites* zone in the lower Hermitage of Lee County may correspond to one near the base of the Salona; the latter formation carries a thick metabentonite near its base, clay *2* and a higher clay *3* which may correspond to the clay at 45 feet in the Hermitage. The Curdsville of southwestern Virginia has been included in the lower Martinsburg formation northeastward.

Chazyan age of the "Murfreesboro"–"Mosheim"–"Lenoir"–"Ottosee" sequence is indicated by its position above the Beekmantown dolomite and below the lower Black River (Pamelia) *Cryptophragmus* beds.

Tentative correlations of middle Ordovician sections mentioned in the text are included in Figure 9.

Acknowledgments.—The field work which served as a basis for this report was done during the summer of 1942. Faunal collections were made and identified by the writer. The problem was suggested by Professor Kay, of Columbia University, who supervised and aided in the study. The writer profited by discussions with Dr. B. N. Cooper, of the Virginia Survey, and Dr. H. N. Coryell, of Columbia University. The manuscript was read and helpful criticisms offered by Dr. W. H. Bucher and Dr. C. H. Behre, Jr. Members of the faculty of Lincoln Memorial University, Harrogate, Tennessee, and of the Department of Geology of the University of Kentucky expressed a friendly interest in the work.

OBSERVATIONS ON PSEUDO-DIKES AND FOLIATED DIKES

WILLIAM J. MILLER
University of California, Los Angeles

ABSTRACT

Criteria for distinguishing between dikes and pseudo-dikes (or xenoliths), particularly when both are foliated, are discussed in this paper. Consideration is given to time and direction of application of stresses which produced the foliation. Examples are mainly from southern California.

INTRODUCTION

When working in regions of crystalline rocks, special care must often be exercised in distinguishing between true dikes and pseudo-dikes (or xenoliths), particularly when both exhibit secondary structure. Thus, certain foliated dikes may easily be mistaken for xenoliths, and various xenoliths closely resemble dikes. Unusually deceiving are certain long, straight, narrow xenoliths, which upon casual examination would be called "dikes." Criteria for making the distinctions are discussed.

Most of the foliated dikes are in or near fault zones in which shearing stresses produce the foliation long after emplacement of the dikes. Some dikes show a primary structure, and still others show no foliation.

Foliation of pseudo-dikes may or may not have resulted from stresses developed during faulting.

KINDS AND AGES OF THE DIKES

The dikes and pseudo-dikes referred to in this paper are of various kinds and ages, most of them being well represented in the San Gabriel Mountains, the crystalline rocks of which are listed in Table 1. The rocks of particular interest in our discussion are Rubio diorite and metadiorite, dioritic facies of anorthosite, fine-grained dioritic dikes, quartz-latite porphyry dikes, lamprophyre dikes, and basaltic dikes. Rocks like most of these occur also in various other parts of southern California.

TABLE 1

CRYSTALLINE ROCKS OF THE SAN GABRIEL MOUNTAINS

Tertiary	Volcanics Basaltic (or diabase) dikes Lamprophyre (or dioritic) dikes Quartz-latite porphyry (or porphyrite) dikes
Late Jurassic	Aplite, pegmatite, and silexite dikes Lowe granodiorite Wilson diorite
	Metasediments; probably late Paleozoic; eastern part only
	Fine-grained dioritic dikes; probably pre-Cambrian
Pre-Cambrian	Pelona schist; metasedimentary; probably late pre-Cambrian Anorthosite; bluish-gray, white, and dioritic facies San Gabriel formation; a complex of Placerita metasediments, Rubio diorite, and Echo granite Echo granite Rubio diorite; hornblende-rich; varies to metadiorite Placerita metasediments; probably early pre-Cambrian

PSEUDO-DIKES

In the course of field work in southern California many examples of pseudo-dikes have been observed by the writer. Eight of these will be discussed to show

the nature of the rocks and the phenomena involved. Two examples from other regions are also discussed.

in the Lowe granodiorite. They are, in reality, pseudo-dikes representing remnants of Rubio diorite cut to pieces by

FIG. 1.—Pseudo-dikes of biotite- and hornblende-rich metadiorite cut to pieces by granodiorite. Upper end of Little Rock Reservoir, San Gabriel Mountains, California.

FIG. 2.—A pseudo-dike of metadiorite (largely biotite schist) with sharp contacts in pegmatitic granite. India Canyon Forest Service road, three-fifths of a mile from summit in western San Gabriel Mountains, California.

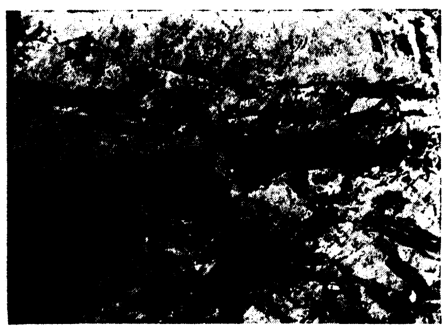


FIG. 3.—Part of the same outcrop shown by Fig. 2, with the metadiorite definitely in the form of xenoliths cut by dikes of the pegmatitic granite.

A very striking illustration of pseudo-dikes occurs near the upper end of Little Rock Reservoir in the northern San Gabriel Mountains. On casual examination this occurrence (Fig. 1) would easily be mistaken for a network of basic dikes the granodiorite. Many granodiorite dikes—often tiny ones—plainly cut the dark, dikelike strips irregularly. The strips vary from massive to highly foliated and sheared. Many large and small xenoliths of the diorite occur in the general vicinity of the reservoir. Near the

dam, several small basic dikes of Tertiary age cut the granodiorite sharply.

By the India Canyon road in the northwestern San Gabriel Mountains, a sharply defined strip (xenolith) of metadiorite, varying to biotite schist (Fig. 2), lies in pegmatitic granite, the latter cutting near-by white anorthosite. Taken by itself, this xenolith would easily be mistaken for a much-sheared basic dike. Close by, in the same outcrop, however, strips of similar material are cut by various dikes of the pegmatitic granite (Fig. 3.)

A narrow, somewhat irregular strip of black metadiorite (Fig. 4), with sharp contacts, lies in light-gray massive granite on the Angeles Crest highway, about a mile below the Red Box summit. Its maximum width is $1\frac{1}{2}$ feet. It is spotted with hornblendes, and it is probably a facies of the Rubio diorite. A tiny dike of the granite cutting across its eastern end proves that it is not a dike. Between this occurrence and Red Box there are a number of similar strips of metadiorite, some cut by granite dikes and others not.

Good examples of interesting pseudodikes are shown in a road cut about a mile east of Lang, near Soledad Sulphur Springs in the western San Gabriel Mountains. They are more or less sheared and foliated, earlier solidified, coarse-grained, dioritic facies of anorthosite in massive white anorthosite. One of these (Fig. 5), about 25 feet in length, is curved, varies greatly in thickness, and thins out at each end. Its xenolithic nature is proved by the fact that locally it is somewhat intimately injected by the anorthosite. Other dikelike xenoliths of

FIG. 4.—A dikelike xenolith of metadiorite with sharp contacts in gray granite cut by a tiny dike of the granite near the right end. Angeles Crest highway, 1 mile down from Red Box summit in western San Gabriel Mountains, California.

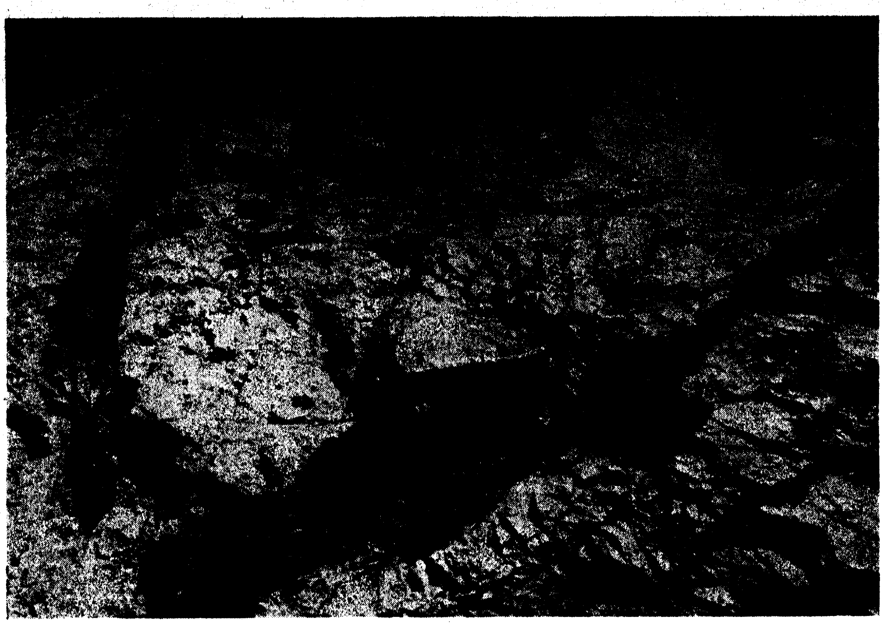


FIG. 5.—A sharply defined, variably sheared and foliated, dikelike xenolith of a dioritic facies of anorthosite in white anorthosite. Foliation is generally parallel to contacts. About 1 mile east of Lang in western San Gabriel Mountains, California.

FIG. 6.—A pseudo-dike of foliated metadiorite in coarse-grained, crudely foliated, bluish-gray anorthosite. Tiny sills of anorthosite cut the metadiorite. Angeles Crest highway southwest of the Monte Cristo Mine in western San Gabriel Mountains, California.

similar material in this vicinity are not cut by the anorthosite.

A dikelike strip or pseudo-dike of hornblende-rich metadiorite in coarse-grained, bluish-gray, crudely foliated anorthosite (Fig. 6) shows on the Angeles Crest highway southwest of the Monte Cristo mine in the San Gabriel Mountains. The anorthosite is somewhat contaminated by dioritic material. Foliation of anorthosite and pseudo-dike are parallel. A few tiny sills of anorthosite cutting the metadiorite prove that the latter is a xenolith.

Northeast of Jacumba, in San Diego and Imperial counties, some of the many large and small inclusions of dark-gray metadiorite in light-gray quartz diorite may, from little distances, easily be mistaken for dikes. A good case in point is the pseudo-dike, 10–15 feet wide, extending across the face of a small mountain 3 miles northeast of Jacumba. Still finer examples occur 1–2 miles west of the mouth of In-Ko-Pah Gorge and about 5 miles west-southwest of Coyote Well. The one studied in detail is exposed for a distance of about 700 yards. It is 20–60 feet wide and remarkably straight (Fig. 7). The dark-gray, variably foliated, biotite- and hornblende-rich, fine to medium-grained pseudo-dike strikes nearly east-west across a low ridge of light-gray, massive quartz diorite and disappears under the desert wash on each side. Contacts against the diorite are generally very sharp. Dips of both pseudo-dike and its generally well-developed foliation are 50°–70° south. The following facts prove that we are here dealing with

FIG. 7.—Part of a remarkable dikelike xenolith of variably foliated, dark-gray metadiorite in light-gray quartz diorite. It disappears under the wash in the middle distance. About 5½ miles a little south of west of Coyote Well, Imperial County, California.

FIG. 8.—Sill-like xenoliths of metagabbro-diorite in foliated, gray, pre-Cambrian granite. Lower Gold Park Canyon, south of Twenty-nine Palms, California.

FIG. 9.—A basaltic dike in quartz monzonite. From a short distance it looks like a xenolith. A pseudo-dike of monzonite extends across it. About 20 miles south of Needles, California.

a xenolith: a few small dikes both of the diorite and of pegmatitic offshoots of it cut the pseudo-dike; and some strips of the metadiorite, about a foot in length, lie in the adjacent diorite, parallel to the pseudo-dike contact. Another similar pseudo-dike occurs about 600 yards farther south of the one just described and parallel to it. A lens of similar metadiorite, 60 feet long and 15 feet wide, lies in the diorite near by.

In lower Gold Park Canyon, south of Twenty-nine Palms, a number of pseudo-dikes of basic material occur as sill-like strips or bands in highly foliated pre-Cambrian granite. These strips have schistose borders about $\frac{1}{4}$ inch wide. The writer, in company with several graduate students, decided, after very careful study, that the dark strips in the highly foliated pre-Cambrian granite shown in Figure 8 are not sill-like intrusives in the granite but, rather, xenoliths of meta-gabbro-diorite in the granite. A study of near-by outcrops shows that these bands are metamorphosed remnants of a large body of gabbro-diorite border, portions of which were cut to pieces by the granite magma. Most of the remnants were largely injected more or less intimately, and even digested, by the granite. In this vicinity several small dikes of Tertiary basalt cut the granite-metagabbro-diorite complex sharply.

A small hill of massive, moderately coarse-grained monozonite, $4\frac{1}{2}$ miles south of Twenty-nine Palms, contains several nearly straight, narrow xenoliths of metadiorite which may easily be mistaken for dikes. These pseudo-dikes range in width to 1 foot and in length to 100 feet. The combination is cut sharply by a small, massive, fine-grained, basic Tertiary dike with chilled borders.

About 20 miles south of Needles numerous large and small dikes of basalt and quartz-latite porphyry cut quartz monzonite. One of the basic dikes (Fig. 9) could, from a little distance, easily be mistaken for a xenolith. It is 35 feet in length, and it ends abruptly at each end. It is fine grained and massive, with chilled borders. A pseudo-dike of the monzonite extends through it.

Two examples of pseudo-dikes outside of California will now be considered. One of these is by the highway 5 miles north of Schroon Lake in the Adirondack

FIG. 10.—A dikelike xenolith of foliated garnetiferous, metagabbro, 18 inches wide, in light-gray, garnetiferous anorthosite. Five miles north of Schroon Lake, Adirondack Mountains, New York.

Mountains, where several sharply defined layers or bands of roughly foliated garnetiferous, amphibolitic material occur in poorly foliated, garnetiferous Whiteface anorthosite. Foliation of amphibolite and anorthosite are parallel. These layers are either highly altered metagabbro or mafic gneiss of Grenville origin. In either case, they are xenoliths; but they could easily be mistaken for dikes—one of them especially (Fig. 10). Close by, in the same outcrop, however, similar mafic bands are cut sharply by the anorthosite. Also, some small shreds of the mafic material have been variably digested by the anorthosite.

In 1938 the writer observed several in-

teresting pseudo-dikes, forming a parallel group in an outcrop near Molde, Norway. These are fine-grained, strongly foliated metadiorite, with sharp, somewhat irregular, boundaries, in foliated, pink, impure, pre-Cambrian granite. Foliation of pseudo-dike and granite are parallel. The smaller pseudo-dikes are seen to pinch out at each end. The largest one, exposed for 40 feet, is 1½–2 feet in width (Fig. 11). A small pegmatite dike, derived from the granite magma, cuts this pseudo-dike very sharply; but the pegmatite is not sharply separated from the granite. Sharp contacts of the metadiorite against both the granite and the pegmatite and the highly foliated and sheared structure of the metadiorite rule out the possibility that we are here dealing with a true dike—even one which could have intruded the granite before final consolidation of the latter. F. F. Grout[1] has described a lamprophyre dike, believed to have cut syenite at Snowbank, Minnesota, before the syenite was entirely crystallized because some syenite cross-cuts the lamprophyre and also because a small pegmatite dike, derived from the syenitic magma, cuts both syenite and lamprophyre. In this case, however, the lamprophyre is massive and its contacts are not sharp.

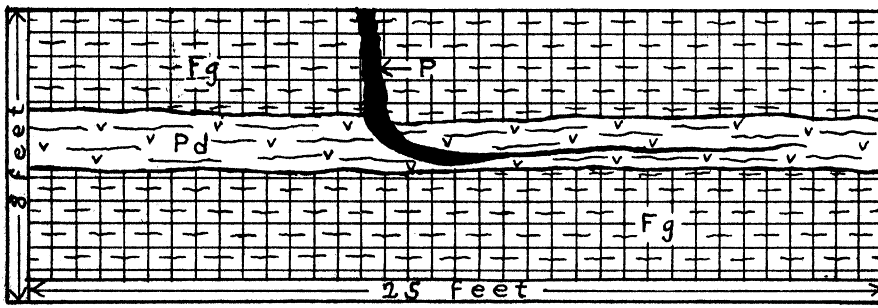


Fig. 11.—Ground-plan sketch of a strongly foliated, basic, pseudo-dike (*Pd*) with sharp contacts in moderately foliated granite (*Fg*). Both are cut by a pegmatite dike (*P*). Near Molde, Norway.

FOLIATED DIKES

In southern California, as elsewhere, many dikes exhibit varying degrees of primary foliation. In this paper, however, examples only of dikes which show secondary foliation are considered.

An interesting network of foliated basaltic dikes shows in a road cut about 1 mile west of Chilao on the Angeles Crest highway in the San Gabriel Mountains. They are of Tertiary age; they cut Lowe granodiorite sharply; and they run in all directions. Part of this network is illustrated by Figure 12, where the maximum dike width is shown to be 2½ feet. Each dike is largely sheared into schist, with foliation always parallel to the walls, no matter how the dikes run.

Figure 13 shows a medium-fine grained dioritic dike 2 feet wide cutting horizontally across pre-Cambrian banded gneiss (on left) and pre-Cambrian diorite (on right) near Barley Flat in the San Gabriel Mountains. The borders of the dike, 2–4 inches wide, have been sheared into a hornblende-biotite schist with foliation paralleling the dike borders.

[1] *Petrography and Petrology* (New York: McGraw-Hill Book Co., 1932), p. 121.

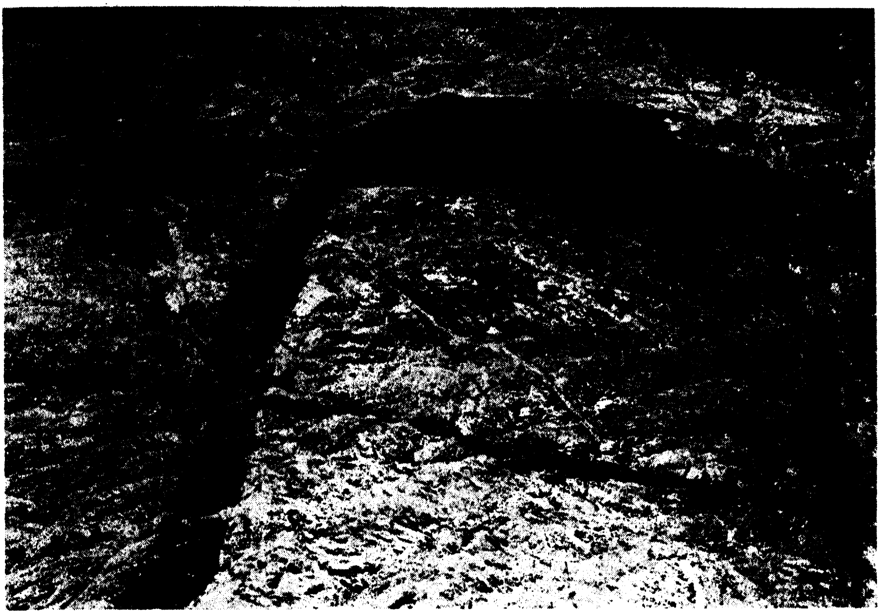


FIG. 12.—Part of a network of schistose basaltic dikes in light-gray granodiorite. Maximum dike width is 2½ feet. Near Chilso in the western San Gabriel Mountains, California.

FIG. 13.—A medium-fine grained, probably pre-Cambrian, dioritic dike, 2 feet wide, cutting horizontally across pre-Cambrian banded gneiss (*on left*) and Rubio diorite (*on right*). Borders of the dike are schistose. Near Barley Flat in the San Gabriel Mountains, California.

Some small sheared masses also occur farther within the dike. This dike is probably pre-Cambrian.

On the Angeles Crest highway in lower Mill Creek Canyon in the middle-western San Gabriel Mountains a fine-grained dioritic dike, $1\frac{1}{2}$ feet wide (probably pre-Cambrian), diagonally and sharply cuts strike and dip, and all are more or less foliated and sheared, always parallel to their borders. In Figure 15 two such dikes are shown sharply cutting white, nonfoliated, but much-fractured, white anorthosite. The small, dark, wedge-shaped mass lying between the two dikes is a xenolith of diorite representing an

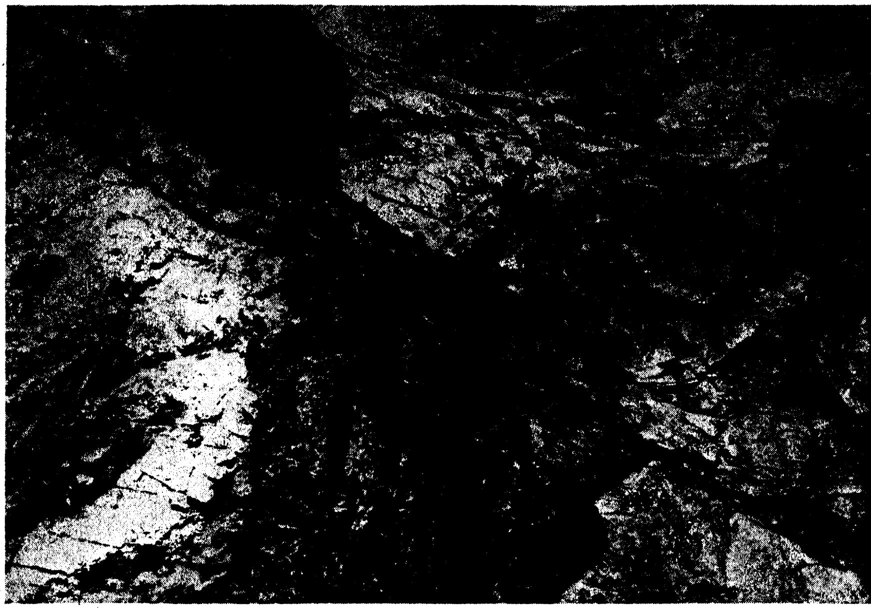


FIG. 14.—A fine-grained dioritic dike, $1\frac{1}{2}$ feet in width, cutting anorthosite with dioritic facies (*on left*). Wilson dioritic (*on right*) cuts both the anorthosite and the dike. Borders of the dike are variably sheared and schistose. Lower Mill Creek Canyon in the middle-western San Gabriel Mountains, California.

massive anorthosite, which varies from nearly pure, bluish-gray to a dioritic facies, as shown in Figure 14. Narrow borders of the dike grade rapidly into schist as a result of shearing. Massive Wilson diorite, on the right in Figure 14, cuts both the dike and the anorthosite sharply.

Along the road in lower Soledad Canyon in the vicinity of Soledad Sulphur Springs, many foliated, basic dikes are excellently exhibited. They vary greatly in earlier crystallized facies of the anorthosite. The steep-dipping, nearly black dike is a medium-fine grained, hornblende- and biotite-rich metadiorite, which is probably pre-Tertiary. Its main, interior portion is only crudely foliated, but its borders are highly schistose and sheared parallel to contacts. It has been offset 3 feet by faulting. A younger, low-dipping, gray, fine-grained, basaltic dike follows the fault and crosscuts the black dike. It is highly sheared, usually schis-

tose, with foliation paralleling contacts. There are some remnants of chilled borders.

Near the outcrop just described, there are two other black, hornblende- and biotite-rich dikes of metadiorite (Fig. 16), similar to the one shown in Figure 15; but they dip steeply in the opposite direction. They vary from 18 to 36 inches in width. Some remnants of chilled borders are preserved in contact re-entrants. One of the dikes exhibits the usual border schistosity parallel to contacts; but its interior is diagonally well foliated and somewhat sheared, thus plainly indicating torsional stress or drag effect. The other dike is variably schistose throughout; but, where thickest, its interior is notably crumpled. Such phenomena have been observed much less conspicuously in a few other basic dikes in the San Gabriel Mountains. This is surprising, in view of the fact that so much secondary schistosity and shearing have been produced in the basic dikes.

One-third of a mile east of Soledad Sulphur Springs, and just east of the road tunnel in the western San Gabriel Mountains, there is a grand display of roughly parallel, basic dikes cutting white anorthosite. Figure 17 illustrates part of this dike system. These dikes are largely nearly black, fine to finer medium-grained, hornblende- and biotite-rich diorite of pre-Jurassic granite age. Dikes and sills of the granite, ranging from tiny to several feet in width, cut the basic dikes sharply. A few small, fine-grained, basaltic dikes of Tertiary age cut the dioritic dikes. Both dioritic and basaltic

FIG. 15.—White, shattered anorthosite is cut by two basic dikes. One is a steep-dipping, black, biotite- and hornblende-rich dioritic dike with schistose borders and offset 3 feet by faulting. The other is a schistose basaltic dike of Tertiary age which follows the fault. East of Lang, near Soledad Sulphur Springs in the western San Gabriel Mountains, California.

dikes vary from massive to highly foliated; and both are locally crumpled, especially their border portions, but some remnants of chilled margins remain. A few small xenoliths or segregation masses of the earlier consolidated, now variably foliated, coarse-grained facies of the anorthosite occur here and there. Four of the rocks just mentioned are shown in detail in Figure 18. This occurrence is a little to the left of the general view illustrated by Figure 17. The white rock is anorthosite. At the lower left is a wedge of the black, foliated, dioritic facies of the anorthosite. All large and small, dark-gray dikes, sharply cutting the anorthosite in the upper one-half of the exposure, are variably foliated diorite. The gray, wedge-shaped mass with the hat on it is late-Jurassic, massive pink granite, sharply cutting the other three docks.

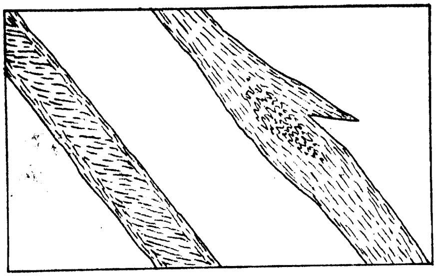


FIG. 16.—Black, fine-grained, schistose, dioritic dikes in medium-grained white anorthosite. One of these, 18 inches in width, shows diagonal, as well as marginal, schistosity; the other is foliated throughout, with crumpling in its thickest portion. Close to the dikes shown in Fig. 15.

Along the road for one-third of a mile

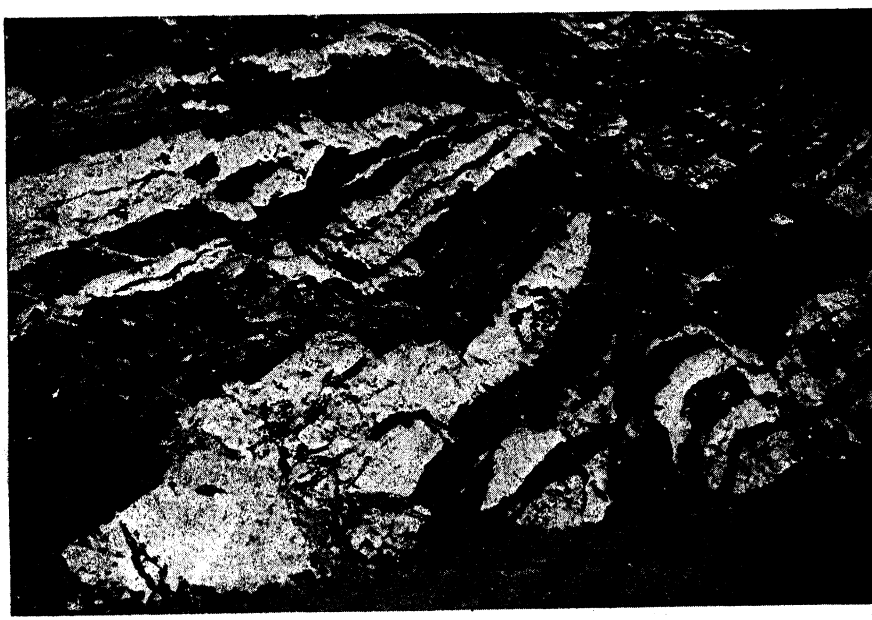


FIG. 17.—Part of a roughly parallel system of black, fine-grained, dioritic dikes cutting white anorthosite. A few small dikes of granite cut both anorthosite and diorite, and a few small basaltic dikes of Tertiary age cut the other three rocks. Both dioritic and basaltic dikes vary in structure from massive to schistose, but neither anorthosite nor granite is foliated. One-third of a mile east of Soledad Sulphur Springs, in the western San Gabriel Mountains, California.

east of the above-mentioned road tunnel in lower Soledad Canyon the main country rock is white anorthosite, cut irregularly and sharply by a good many of the dark-gray, fine-grained, biotitic diorite dikes. The larger ones have schistose borders. The anorthosite is cut very irregularly by much pink granite and pegmatite, small sills and dikes of which cut the basic dikes.

FIG. 18.—White anorthosite with a black, wedge-shaped xenolith of an earlier foliated, dioritic facies (*lower left*), cut irregularly by a system of black dioritic dikes with variable foliation parallel to contacts. A wedge of massive pink granite (*hat on it*) cuts the other three rocks. Just west of the more general view shown in Fig. 17.

An interesting case of two strongly foliated basic dikes in marble in the southern Sierra Nevada has been described by Cordell Durrell.[2] The foliation of each dike is parallel to the dike walls, while that of the country rock is nearly parallel to the plane of the picture (Fig. 19). Thus there are three directions of foliation mutually at right angles to each other. Durrell says that these dikes were intruded before metamorphism of the marble.

In various parts of southern California the writer has found quartz-latite porphyry dikes of Middle Tertiary age. They occur singly and in groups. They vary in width from a few feet to 50 feet. Contacts are sharp, and dips are steep. All have fine-grained groundmass, and nearly all are very massive. Exceptionally, some of these dikes are well foliated. This is true of a group of roughly parallel quartz-latite porphyry dikes in the Chocolate Mountains about 10 miles northeast of Niland. They cut quartz

[2] "Metamorphism in the Southern Sierra Nevada Northeast of Visalia, California," *Univ. Calif. Pub., Dept. Geol. Sci.*, Vol. XXV (1940), p. 113, Fig. 28.

monzonite sharply. One of these is shown by Figure 20. They are strongly sheared and are foliated parallel to their walls, Victorville region. The foliation is usually accentuated by crude bands of biotite or chlorite.

Fig. 19.—Two strongly foliated, basic dikes nearly at right angles to schistosity parallel to dike walls. Country rock is marble. Near Lemon Cove in the Southern Sierra Nevada, California. Photo by C. Durrell.

with a dip of about 40°. A similar dike, strongly sheared and foliated, occurs near the mouth of Monumental Canyon southwest of Needles, and another 4½ miles east of Sidewinder Well in the

SIGNIFICANCE OF THE FOLIATION

By way of explanation of the secondary foliation in the above-described dikes and of similar foliation in some of the pseudo-dikes, several factors should be

considered. It is probable that, in some measure, secondary foliation has been impressed upon, and accentuated by, primary foliation. In view of the fact, however, that many dikes of similar composition, but without primary foliation, occur in the same general regions above mentioned, such primary foliation may be ruled out as an important contributing factor.

Composition is an important factor. By far, most of the strongly foliated dikes above described are dioritic or even more basic, and they are rich in biotite and hornblende—minerals which favor development of foliation, even schistosity, under shearing-stress conditions. Few, if any, of the quartz-latite porphyry dikes and the still more numerous aplite dikes in the San Gabriel Mountains exhibit any secondary foliation. In many cases, however, they are much shattered. The few observed cases of strongly foliated quartz-latite porphyry dikes in other parts of southern California involve rocks containing several per cent of biotite or hornblende in a very fine-grained groundmass.

Most important in producing the secondary foliation of the dikes has been dynamic metamorphism under relatively shallow-zone conditions within the earth. The finest and most numerous examples of such foliated dikes observed by the writer in southern California occur in the San Gabriel Mountains. This whole mountain block of 1,200 square miles of crystalline rocks has been squeezed up thousands of feet between steep indipping faults during Quaternary time. During this time the whole block has shifted northwestward at least several miles. Thus there developed a strong torsional stress within the mountain block, resulting in a great number of large and small faults running in various directions. The writer has never seen another rock mass of comparable size so remarkably fractured and shattered, particularly along the more important fault zones. Conditions were exceptionally favorable for the development of shearing stresses. All, or nearly all, of the more strongly foliated dikes are in or near important fault zones, developed, and accompanied by more or less shearing, long after emplacement of the dikes—for example, Tertiary and older dikes foliated

FIG. 20.—Sheared and foliated quartz-latite porphyry dike in quartz monzonite. Foliation is parallel to walls of dike. Ten miles northeast of Niland in the Chocolate Mountains of southern California.

by Quaternary stresses. It is remarkable how effective components of stress have often been so distributed as to produce shearing and foliation in the basic dikes with widely variable trends even in single outcrops, as so strikingly exhibited in lower Soledad Canyon. Evidently, these basic, biotite- and hornblende-rich dikes in or near fault zones were very susceptible to shearing parallel to their walls, no matter what their strike and dip. In this connection, it is important to note that the adjacent country rocks—usually medium to moderately coarse-grained anorthosite or quartz monzonite—were much shattered but not foliated. All this indicates that the dike foliation developed under relatively shallow condi-

tions and that the amount of movement during the shearing of any dike was not more than a few inches.

The writer has observed many basic dikes, particularly those of Tertiary age, which are not foliated in the San Gabriel Mountains. These are largely in localities well away from prominent Quaternary fault zones, but some (e.g., West Fork San Gabriel Canyon) occur in older fault zones in which little or no movement has occurred since the dikes were emplaced.

In the case of the above-mentioned quartz-latite porphyry dikes—much poorer in minerals favorable to development of foliation—only those striking parallel to near-by important fault zones became secondarily foliated. In these dikes foliation paralleling the faults was produced. This is conspicuously true of the group of such dikes close to the San Andreas fault zone 10 miles northeast of Niland, where the dikes are strongly sheared and foliated and the country rock of quartz monzonite is badly shattered.

The already mentioned foliated, basic dikes occurring near Lemon Cove in the Sierra Nevada do not seem to be associated with a fault. They are regarded by Durrell as premetamorphic dikes whose secondary foliation in different directions was produced by components of stress at the time of folding and metamorphism of the country rock, which is marble. Probably in this case the relatively weak marble yielded to stress by folding and flowage, while the more resistant basic dike material yielded to components of the stress by shearing.

Some of the xenolithic pseudo-dikes, under favorable conditions of mineral composition and position with respect to fault stresses, were secondarily foliated in a manner similar to that of the foliated dikes. A good example is shown by Figure 5. In most of these cases, however, foliation is believed to have developed before or during the time of intrusion of the enclosing country rocks, with possibly some later accentuation of foliation.

According to P. Eskola,[3] narrow dikes of amphibolite with "parallel metamorphic structure" and with no definite strike occur in various parts of southern Finland. He describes several, stating that they are dioritic and basaltic in composition. And he says that "to explain the origin of the foliation it is not necessary to assume orogenic movement and stress in more than one direction"; that hornblende could result from the metamorphism; and that, in "accordance with the principle of Riecke, the individuals of this mineral, when crystallizing under stress, would assume a parallel arrangement" in the direction of the strike of the dikes. With this explanation the writer is in partial agreement; but, in the cases described in the present paper, a considerable component of stress has also often produced notable shearing in the dikes parallel to their walls. Also, instead of a general regional pressure, stresses, developed in or near important fault zones, caused the foliation of the southern California dikes. This may explain why groups of dikes, as well as individual dikes in southern California, exhibit such striking local variations in degree of foliation, as compared to the greater uniformity in those of Finland. Eskola believes that recrystallization of original dike material was very important; but the present writer believes that foliation of the southern California dikes has resulted from dynamic metamorphism, involving little, if any, recrystallization.

[3] "Petrology of the Orijärvi Region in Southwestern Finland," *Bull. Comm. Geol. de Finlande*, No. 40 (1914), pp. 113–18.

THE PHYSIOGRAPHIC HISTORY OF AN EOCENE SKYLINE MORAINE IN WESTERN MONTANA

WALLACE W. ATWOOD AND WALLACE W. ATWOOD, JR.
Clark University, Worcester, Massachusetts

ABSTRACT

Gravelly Range in southwestern Montana has long attracted attention because of the huge amount of loose, water-worn material along its crest. Our studies in that range have demonstrated that the material along the skyline, in large part at least, is a glacial moraine. Some of the finer material may be outwash associated with glaciers. This moraine was deposited by alpine glaciers which descended from mountains that no longer exist and was placed in a valley during Eocene time. There it remained through the long periods of mid-Tertiary erosion when the widespread Rocky Mountain peneplain was developed. Today, owing to the subsequent uplift and erosion of the range, it is the high place or crest of a mountain range.

INTRODUCTION

On the crest of a subdued mountain area, between the Madison and Jefferson valleys and extending southward from Virginia City, there is a huge amount of bouldery and gravelly material which has attracted the attention and interest of many prospectors and students of geology. The highland mass upon which this loose or uncemented material rests is known as the Gravelly Range. To the northward are the Tobacco Root Mountains. To the west and about parallel to the Gravelly Range are the Green Horn and Snowy Crest Mountains. Eastward, across a broad valley lowland, are the rugged peaks of the Madison Range (see Fig. 1).

There are those who believe that the rich placer deposits of Alder Gulch, west of Virginia City, must have come down the slopes of the Gravelly Range. Many have prospected over and over again in the crest-line accumulation of gravelly materials. There they have found gold. Perhaps there is a relationship between these two deposits, but that can best be suggested after we have presented an analysis of the problems involved in a study of the physical history of this skyline moraine. The story to be told reflects the history of the entire mountain area during the last fifty to sixty million years.

In the summer of 1941, while we were on a brief reconnaissance visit in this part of Montana, the high-level bouldery deposits near the southern end of Gravelly Range impressed both of us as of some very special significance, and we naturally formulated working hypotheses that needed to be tested. In July, 1944, the senior author revisited the area.

THE MORAINIC MATERIAL

The largest of the boulders in this deposit, 10–12 feet in diameter, are granites, gneisses, or schists. Associated with the coarsely crystalline rocks are immense quantities of quartzites and fine-grained, dark-colored igneous rocks with pebbles and cobblestones of quartz, jasper, porphyry, and conglomerate. A few boulders were found that are presumably from a pre-Cambrian conglomerate, containing quartz pebbles and many pieces of jasper that are well worn. This conglomerate is a very hard, compact, and ancient-looking rock, like many of the pre-Cambrian conglomerates of the western mountain areas. On some of the fine-grained, black igneous pebbles of the gravelly deposit, glacial striae are still

preserved. This aggregation of crystalline and metamorphic materials suggests a pre-Cambrian complex as a source.

The skyline moraine rests upon Cretaceous sandstone and shale. The underlying formations dip gently to the westward, and the harder strata have caused the development of a series of cuestas with east-facing escarpments in which the different strata in the sedimentary rocks are well exposed.

The best and most striking examples of striation and grooving are on certain of the huge gneisses and granite masses. The one shown in Figure 2 is a gneissic boulder, with the color bands running transversely to the lines of grooves and scratches. The surface of this boulder is now somewhat pitted, for it may have been exposed for many years to the processes of weathering. Numerous other boulders showed evidence of intense glacial rubbing, polishing, planation, and striation.

Much of this bouldery mass is wonderfully well exposed in west-facing escarpments along the crest of the range south of Black Butte. At this place the escarpment is due to the preserving effect of the morainic capping. In the moraine there is a large amount of material so beautifully water worn that it gives the impression, when first examined, that it must be a stream deposit. By far the greater part of this moraine is water worn, but in the mass there are many subangular rocks that show very little water wear. The deposit shows no signs of stratification or even the rough assortment of materials commonly present in stream-deposited sediments. In Figure 3 it is clear that the material is not assorted.

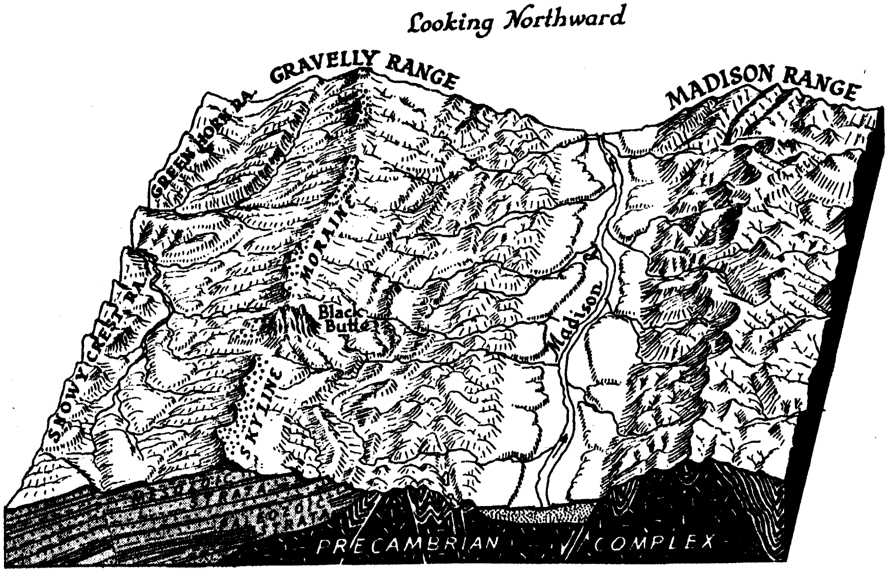


FIG. 1.—A simplified structural section and relief sketch. There is no accurate map available on a scale which would carry as much detail as is shown here.

The large masses of morainic material

FIG. 2.—A glaciated gneissic boulder about 10 feet long, located near the crest of Gravelly Range. The gneissic texture is at right angles to the glacial striae and grooves.

FIG. 3.—An exposure in the skyline moraine on the crest of Gravelly Range about three miles south of Black Butte.

now present are presumably but remnants of a much more extensive glacial deposit. The moraine is 50–100 feet thick at several localities at or near the crest of the range. One of the best exposures extends for about 2 miles along the crest of the range south of Black Butte. Other large remnants are north of Black Butte.

SOURCE OF THE SKYLINE MORAINE

In 1941 we found evidence which indicated that this crest-line material must be of glacial origin; but we were puzzled as all students of this problem must have been puzzled, in finding what seemed to be a moraine along the divide of a mountain range. Where could the materials have come from? There are no cirques or U-shaped canyons through which it could have moved to this location. There is no higher place in the present topography from which this material could have come. Black Butte (10,546 feet), the only higher rock mass near the crest of the range, is composed of extrusive basalt, but it did not contribute to the materials in this moraine. Furthermore, it shows no signs of glaciation. We are certainly not dealing with a Pleistocene moraine.

Since most of the very large striated boulders in this skyline moraine are crystalline and metamorphic, they must have been moved by ice from outcrops 10–15 miles away, where such rocks occur. The conclusion is unavoidable that the entire mass must have moved downhill, and therefore we must imagine remarkable changes that have taken place in this region since the moraine was deposited. The crest line of Gravelly Range, where the moraine now rests, must have been, at some time in the past, the low place or valley bottom in an ancient topography. Where there is blue sky today, there were lofty mountains with perennial snow fields, which gave rise to alpine glaciers. The present crest of the range was the floor of a mountain canyon, through which a great mass of debris-laden ice moved. That must have been in Eocene time when lofty mountains with alpine glaciers existed in various parts of the Cordilleran province of North America. A mountain landscape in which there were many glaciers has disappeared. An old valley floor has become the high-line divide of a mountain range of a later generation.

During field studies conducted in July, 1944, the locations of the Black Butte tillite were visited, the character and distribution of the crest-line moraine were rechecked, and the areas of pre-Cambrian rocks near the southeastern margin of Gravelly Range and in the neighboring mountain area farther to the east were visited. There is no suitable topographic map published for the Gravelly Range area, but the heaviest and most continuous masses of the crest-line moraine are located on sketch maps drawn in the field (see Fig. 1).

Gravelly Range is not a rugged mountain area. The mountain tops have been very much softened and subdued by erosion, and there is abundant evidence of an old-age topography in the summit area. Viewed from the Madison Valley to the east, the skyline of Gravelly suggests a peneplain surface. When seen from good outlook points in the range, the peneplain can be easily reconstructed in imagination; but it is evident that this old-age erosion surface has been uplifted and dissected. The mountain forms of today are due chiefly to the dissection of a peneplain. Our studies throughout the Rocky Mountain region lead us to believe that many of the surfaces in this range are remnants of the widespread Rocky Mountain peneplain developed during mid-Tertiary time.

This area has been peneplained, uplifted, and dissected since the moraine was deposited; and its modern valleys contain quantities of boulders, presumably derived from the moraine. A considerable number of large granite boulders were found in the headwaters of Gazelle and Barnett creeks. Barnett contains the best examples of striated boulders. They can be seen today on their downward journey from the moraine on the divide between the east- and west-flowing streams of Gravelly Range.

OTHER EOCENE MORAINES

Harold W. Scott reports finding glacial material underneath the Black Butte basalts and concludes that the deposit is of Eocene age.[1] Mr. Scott reports that he has found well-striated glacial stones in the deposit near the southwest margin of Black Butte and, overlying the tillite, an andesitic tuff, which is, in turn, overlain by the Black Butte basalt. The glacial deposit, which he calls the "Black Butte till," rests upon Cretaceous sandstones and shales at that locality, and it is undoubtedly part of the system of moraines reported in this article. Because of the geologic sequence, Mr. Scott is perfectly justified in correlating the deposit which he discovered with the Ridgway tillite discovered by the senior author of this paper in 1915.[2] He correlates it also with a similar deposit discovered in British Columbia by C. W. Drysdale.[3] He also relates his discovery to Eocene deposits of glacial till in Colorado reported by the two authors of the present article.[4] Mr. Scott reports the deposit of till as varying in thickness "from a thin veneer to as much as 200 feet." He refers to Black Butte and the associated basalts and tuffs as of mid-Tertiary age.

THE PHYSIOGRAPHIC HISTORY

In reviewing the physiographic history of the skyline moraine on the Gravelly Range we must think back to the time when the first uplifted masses of the Rockies were being carved into bold, rugged features with large catchment basins in which snows accumulated. The Rocky Mountain ranges were all brought into existence in the great Laramide revolution at the opening of Tertiary time. As uplift proceeded, weathering and stream erosion began in each range; and, as time passed, bold majestic mountain forms may have characterized the topography of the Eocene landscape.

In those ancient mountains that no longer exist the Ridgway tillite and the other Eocene tills of about the same age were deposited. They have been discovered in various parts of the Cordilleran section of the continent and undoubtedly will be reported from many other localities in the western mountains of the United States and Canada. In or near the area of the Gravelly Range there were mountains composed of the pre-Cambrian complex of granites, gneisses, schists, quartzites, conglomerates, and fine-grained igneous rocks. Other mountains made of sandstones, limestones, and shales were located west of the pre-Cambrian rocks above the present Gravelly Range. They were carved out of the bedded formations that rested on the slopes of the ancient core-rocks (see Fig. 1).

In the mountains of Eocene time,

[1] "Eocene Glaciation in Southwestern Montana," *Jour. Geol.*, Vol. XLV (1938), pp. 628–36.

[2] "Eocene Glacial Deposits in Southwestern Colorado," *U.S. Geol. Soc. Prof. Paper 96B* (1915).

[3] "Geology of Franklin Mining Camp, British Columbia," *Geol. Surv. Canada, Mem. 56* (1915).

[4] "The Gunnison Tillite," *Jour. Geol.*, Vol. XXXIV (1926), pp. 612–22.

snows accumulated and formed alpine glaciers, which descended into the valleys of that ancient mountain area. The crest line of the Gravelly Range of today was, at that time, the low place—the place toward which the streams flowed, toward which the ice moved. It was the place where the debris of streams and of the ice of that day was deposited.

During the long mid-Tertiary period, throughout the Rocky Mountain region erosion was going on in a vigorous way, and in the great orographic lowlands between the mountain ranges vast quantities of material accumulated. That was a period of erosion and of filling. As presented by the present authors in 1938,[5] this was the time when a widespread cycle-end surface was developed. In most of the mountain ranges thus far examined, extensive areas were reduced to the peneplain stage. In many instances a chain of crest-line peaks rose above that peneplain as monadnocks. The Snowy Range in the Medicine Bow Mountains is a good example. Another excellent example is seen in the crest-line peaks of the Wind River Range. Pike's Peak, Long's Peak, and other crest-line peaks in the Front Ranges of Colorado are monadnocks, remnants of mountains that were formerly thousands of feet higher. Mount Uncompahgre, Mount Sneffels, and the Needle Mountains, in the San Juan area of southwestern Colorado, are other examples of monadnocks rising above that erosion surface which developed near the close of the mid-Tertiary period of degradation and aggradation.

The cycle-end surface was produced during the millions of years that elapsed from late Eocene time to about the opening of the Pliocene epoch. That was an era when stream erosion continued with but local interruptions in the mountain ranges of the Rocky Mountains. The erosion surfaces which developed in the different ranges at this time are the most widespread of all such surfaces in the Rocky Mountain region, and we prefer to group them together and refer to this composite erosion surface as the "Rocky Mountain peneplain."

It was during this same long period of mid-Tertiary erosion that a vast amount of valley-filling was taking place. Huge quantities of fine material accumulated in the intermountain lowlands and on the great plains east of the Rocky Mountains. In the basin region of Wyoming the alluvial waste actually buried the granite ranges of that area. It was then that the Bozeman beds accumulated in many of the valleys of western Montana. Those beds contain vast quantities of volcanic ash, and some are beds of lacustrine origin.

In many parts of the Rocky Mountain region vulcanism played a large part in the activities of mid-Tertiary time. The lava plateau of the San Juan Mountains was built up during this period. The Absaroka plateau of volcanic materials came into existence, and the vulcanism of Yellowstone National Park dates from this same period of time. In the Crazy Mountains and at many other places lavas were poured out on the surface of the country. In the Gravelly Range, Black Butte and associated flows of lava, which appear to have come through fissures, are of mid-Tertiary age.

The long cycle of erosion, during which the peneplain surfaces were developed and the valley lowlands largely filled with debris, was closed in about mid-Pliocene time by a general uplift of the whole Rocky Mountain region. The

[5] "A Working Hypothesis for the Physiographic History of the Rocky Mountain Region," *Bull. Geol. Soc. Amer.*, Vol. XLIX (1938), pp. 957–80.

mountain ranges, where movement had been started at the opening of Tertiary time, were somewhat more uplifted than the intermontane lowlands. This emphasized the mountains and soon gave increased relief to the general landscape. At many places faulting occurred at the margins of the ranges, and in some cases the movement is measured by thousands of feet. Faulting helped to define many of the ranges in western Montana. At the east margin of the Gravelly Range, near the south end, there is evidence of recent faulting. At the west base of the Madison Range there is abundant evidence of slipping.

In the immediate vicinity of the Gravelly Range the Madison and Jefferson valleys were filled, nearly if not quite, to the peneplain level with alluvium that came from the adjoining mountains as the erosion of the highlands proceeded. The situation in the Madison Valley is of special interest, for that valley still contains hundreds of feet of alluvial wash from the neighboring ranges. Some of it is deltaic in structure, which means that a lake existed in this valley for a time. It is a bouldery, gravelly deposit in the bench lands used by the ranchmen for crops or as pastures. At the north the broad lowland of Madison Valley comes suddenly to an end at the Norris Hills. There the present Madison River plunges into a gorge, which it follows for several miles. The river is an antecedent stream and one that was superimposed. Perhaps the mid-Tertiary alluvial filling of this area entirely covered the Norris Hills; perhaps it rose 1,000 feet on the sides of the Madison and Gravelly ranges that border the present Madison Valley depression. The river flowing at that higher level found its way northward to join the Jefferson and the Gallatin near the present location of Three Forks to make the Missouri River.

When the Madison River, flowing on a mid-Tertiary alluvial surface, was rejuvenated during the Pliocene uplift, it began at once to lower its channel through the alluvium into the hard rocks beneath. In time it became superimposed upon the hard rocks and began cutting the narrow V-shaped gorge east of Norris. It has cut a valley through the Norris Hills and excavated the modern valley of the Madison Valley, upstream from the Norris gorge. At the south end of the Norris gorge the stream has been ponded, and a reservoir is maintained for electric power. If the entire gorge was filled, much of the Madison Valley lowland soon would be flooded.

Similar stories are associated with many of the valleys of western Montana; for most of the streams have been superimposed upon mountain ridges which were buried during the period of mid-Tertiary filling.

By the close of the long cycle of erosion and deposition of mid-Tertiary time, the summit of Gravelly Range must have had very little relief. The main crest and many of the secondary cuesta forms in their summits appear to be parts of the Rocky Mountain peneplain. The modern valleys all date from the uplift of that surface and are clearly due to the dissection of the peneplain.

Black Butte now rises to an elevation of 10,546 feet and about 1,500 feet above its immediate surroundings, for the crest line of the range is at about 9,000 feet. It may well have been a much more conspicuous peak than it is today. The belt of pre-Cambrian rocks at the southeast margin of the range has given rise to an area of softened hills, in which the schistose character of much of the rock has produced an alignment of some of

the elevations. The granite and gneissic masses have produced other rounded hills or low mountains. The harder strata in the inclined sedimentaries that constitute the major portion of Gravelly Range have produced the cuestas that may have remained, even during the peneplain stage. They are conspicuous features in the central portion of the range today. Lava caps like that on Lion Mountain and Flatiron Mountain have preserved those elevations a little above the general level of the peneplain.

The skyline moraine, which is the center of interest in this article, may have stood for a time a little higher than some of the hills of sandstones and shales, for gravels and boulders associated with sands often serve as a preserving cap in the landscape. Rain waters sink into such deposits and therefore do not erode them so rapidly as they do a cemented rock formation. Boulder caps and gravel caps often have the effect of preserving the summit areas of monadnocks.

THIS MORAINE AND THE ALDER CREEK GRAVELS

The general uplift that came during Pliocene time rejuvenated all streams in the Rocky Mountain region and resulted in the dissection of the cycle-end surface developed during mid-Tertiary time. The development of all or nearly all the valleys in that part of the continent dates from that uplift. No positive evidence of Pleistocene glaciation was noted in the portion of the range which we have examined. There are many examples of landslides, solafluction, and névé work but no modern glacial cirques, no modern U-shaped glaciated valleys, and no recent moraines.

Perhaps this story of Gravelly Range throws a little light on the intricate history of Alder Gulch, where millions of dollars' worth of gold have been recovered. Alder Gulch is one of the valleys cut into the mountain area since the Pliocene uplift, that is, after the close of that long mid-Tertiary period of erosion and deposition and long after Eocene glaciation. As the range was dissected, whatever material there was available was washed into the modern valleys and the re-washing of gravels that were on the summit of the range might well have led to a greater concentration of the gold content of such gravels.

Since gold has been recovered in some of the gravels associated with the summit of the range and there are many claims staked out along that crest line, it is interesting to imagine the streams and glaciers of the Eocene period of erosion, working when mountains existed that are no longer present in the region, having done the first work of concentrating gold in the stream bottoms in this part of the West. The gold which accumulated in Alder Gulch may first have been brought to the summit areas of the mountains and in the later cycle of erosion been concentrated in the younger valleys.

The studies here reported were not undertaken with the purpose of throwing any light on the history of the gold-bearing gravels near Virginia City; but no one can visit that region and show any interest in the mountain area without being asked many times where those gold gravels came from. There are prospectors and others today who would mortgage their homes to discover the answer to that question. Perhaps these studies may lead to a further examination of the area with the idea of throwing more definite light on the stages in the history of the gold which has concentrated in Alder Gulch.

THIS MORAINE AND THE PHYSICAL HISTORY OF THE ROCKY MOUNTAINS

The interpretation herein presented of the skyline moraine on the Gravelly Range fits perfectly into the general hypothesis which we have reported for the physical history of the Rocky Mountains. No one should now be surprised at finding gently rolling erosion surfaces near or at the summit of a range. The chances are that they are parts of that widespread Rocky Mountain peneplain that has been described and defined over and over again. No one should be surprised to find almost anywhere in the Rocky Mountains remnants of the Eocene glacial moraines. They were undoubtedly widespread and may have been very abundant during that glacial age and for a long time after that period.

The unique situation which we are reporting here is a moraine that must be of Eocene age which is not covered by any mid-Tertiary or younger formations. It rests on the summit of a range. As far as we can determine, it never has been covered, but during the physiographic history of the area it has gone through a series of interesting changes in its topographic relationships. When deposited, it was in the low place of a mountain area. Those mountains were worn away, and it came to be a part of a widespread surface of erosion. Then that surface was uplifted and dissected, but as yet it has been left, and it serves as the high place and the drainage divide in the range of today. This moraine has rested where it now lies for 50 or 60 million years. Since the late Pliocene uplift it has come to be a crest-line moraine on a mountain range. It will not remain much longer.

STREAM SELECTIVITY IN THE MIDDLE APPALACHIAN VALLEY

HENRY S. SHARP
Barnard College, Columbia University, New York City

ABSTRACT

Cambro-Ordovician limestones and Ordovician shales largely underlie the Great Appalachian Valley between Virginia and Pennsylvania. By differential erosion the limestone areas have been lowered considerably below the level of the shales; but, in spite of this obvious weakness of the limestones, Conococheague Creek and other streams show remarkable adjustment to shale belts. These courses are interpreted as subsequent and as due to the greater weakness of the shales toward concentrated stream erosion, whereas the limestones have been dissolved by unconcentrated rainwash. No change in climate seems necessary to explain this stream selectivity.

Between Chambersburg, Pennsylvania, on the north and the Potomac River on the south the most important stream draining the Great Valley is Conococheague Creek. This part of the Great Valley, as is generally true for its entire length, is developed on Cambro-Ordovician limestones and Ordovician shales, whose lesser resistance to erosion, compared with the older rocks on the east underlying the Blue Ridge province and the younger conglomerates and quartzites making the Appalachian ridges, is responsible for the lower elevations of the Valley. Structurally, the Valley rocks are continuously but openly folded and in many places faulted,[1] with the structure determining the major topographic trends, here somewhat east of north.

Two major topographic levels characterize the Great Valley in the Chambersburg region; the higher of these, occurring at elevations of 700–750 feet on hills of Martinsburg shale west of Chambersburg, descends to elevations of 550 feet on hilltops of the same shale near the Potomac River. This surface is generally known as the Harrisburg peneplain; and it is, to a very remarkable degree, now preserved only in the Ordovician shale areas. In passing from the Martinsburg areas to those underlain by the Cambro-Ordovician limestones, one commonly drops down over a steep grade to a surface lying between 100 and 200 feet below the Harrisburg surface on the shale. This lower surface, restricted to the limestone, has been termed the "Somerville peneplain." It is best developed on the purer limestones, whereas impure and cherty limestones may rise above it toward the Harrisburg level. This two-story character of the Great Valley is well developed from New Jersey southward at least to Virginia.

The occurrence of these two levels is usually explained by the statement that, following the Schooley cycle of peneplanation, evidences of which remain on top of the Blue Ridge and the Appalachian ridges, rejuvenation caused the erosion of the Harrisburg peneplain across the limestone and shale areas of the Great Valley indiscriminately. Uplift then occurring in the resulting partial cycle, time was allowed for local peneplanation of only the weaker-rock areas, and hence the limestone alone was reduced to the Somerville level. This his-

[1] G. W. Stose, "Mercersburg-Chambersburg Folio," *Geol. Atlas of the United States, Fol. No. 170* (Washington, 1909).

tory demands that the Valley limestones be less resistant than the shale toward processes of denudation. Geologic maps of the Mercersburg and Chambersburg, Pennsylvania, quadrangles show, however, that to a very notable extent the larger streams are located upon the Martinsburg shale, presumably the more resistant rock. From Chambersburg to the Potomac, a distance of about 25 miles, the Conococheague Creek confines its rambling and entrenched meandering course largely to the shale; and the West Branch of the Conococheague, flowing eastward out of the Appalachian ridges, turns southward to follow a shale outcrop for 6 miles between limestones on each side before it cuts eastward across one limestone belt to join the Conococheague on another shale belt. Thus the problem is to explain why the Conococheague follows the shale in an area where the lower regional elevation of the limestones suggests that streams should be subsequent upon them.

Still more striking stream adjustment to the shale is seen along the very meandering lower course of Conococheague Creek between the Pennsylvania boundary and the Potomac River. This region is shown on the Williamsport, Maryland, quadrangle; and the course of the Conococheague has been figured[2] and briefly discussed in *Guidebook 7, 16th International Geological Congress.* As shown in Figure 1, Conococheague Creek, flow-

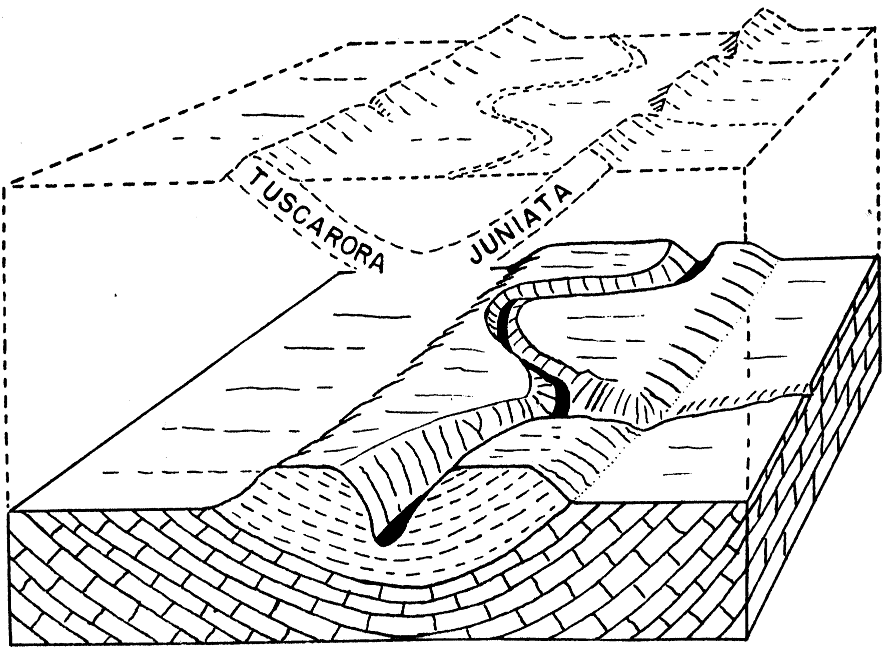


FIG. 1.—Diagram showing synclinal position of Conococheague Creek on shale belt. Top part of diagram shows supposed position of creek in Schooley time, according to hypothesis of inheritance set forth in *Guidebook 7*. (Modified from *Guidebook 7, 16th Internat. Geol. Cong.*).

[2] D. Johnson, F. Bascom, and H. S. Sharp, "Geomorphology of the Central Appalachians," *Guidebook 7, 16th Internat. Geol. Cong.* (Washington, 1932). See p. 45 and Fig. 26.

ing in an entrenched meandering course, is confined to a belt of shale with the older limestone on each side. The shale belt is about 2½ miles wide and occurs along

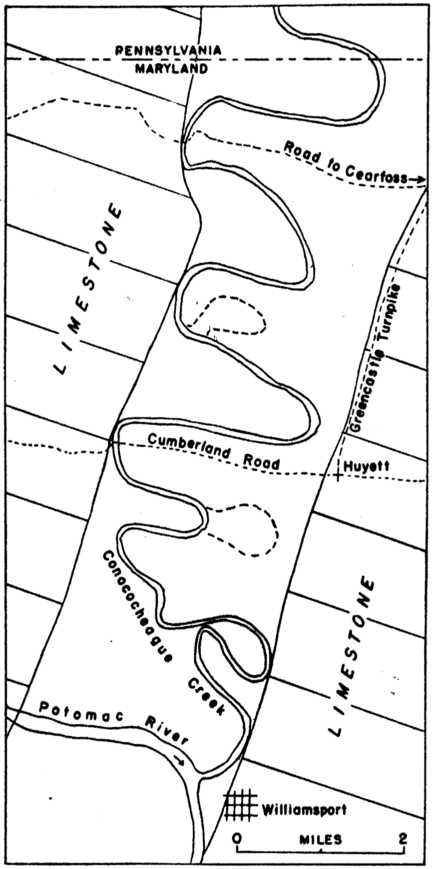


FIG. 2.—Map showing course of Conococheague Creek between Pennsylvania boundary and Potomac River with respect to limestone and shale areas. Heavy broken lines show abandoned high-level meanders.

the axis of the Massanutten syncline. Broad hilltop remnants on the shale at elevations around 560 feet apparently preserve the Harrisburg peneplain and form a broad, dissected, synclinal upland standing above the limestone areas to east and west. Considerable areas underlain by the limestone are less than 500 feet in elevation; and an observer on the limestone looking toward the shale upland from either direction, but particularly from the east, finds the contrast in elevation very conspicuous. From Cearfoss, Maryland, south nearly to Williamsport, a distance of 7 miles, the contrast is truly remarkable; and one standing on the limestone along the Greencastle Turnpike anywhere within this distance and looking westward sees a steep and continuous slope, almost an escarpment, rising sharply, sometimes as much as 100 feet or more above him. This slope marks the eastern contact between the shale and the limestone. The western contact is not marked by a continuous escarpment, but every road passing from upland remnants on the shale to the limestone passes from elevations above 500 feet to those under. Within the shale belt thus topographically defined, Conococheague Creek, entrenched 150–200 feet, makes eleven meander bends (Fig. 2). In addition, the map shows at least two abandoned high-level meanders, so that in a distance of 8 miles, between the Pennsylvania boundary and the Potomac River, the Creek has had a total of thirteen meanders. In this distance its present length is about 25 miles, and with one possible exception it is confined entirely to the shale. Most meanders lie wholly within the shale belt and do not reach the contact. But the meander crossed by the Cumberland Road west of Huyett shows limestone on the outside but not on the inside of the curve, as does also the next westward meander to the north. Here the suggestion is strong that the stream swung outward on the undercut slope until it reached the limestone, which has retarded further lateral migration.

A possible transgression onto the lime-

stone occurs on the westward-swinging meander west of Cearfoss.[3] Inspection by the writer disclosed limestone in the western part of the stream channel and the undercut slope at this point, although the contact was not located. Thus, along the Conococheague there is a record of very remarkable, perhaps perfect, adjustment, made more impressive when it is considered that this is the only belt of shale in this part of the Great Valley and that it is paralleled by belts of limestone several times its width to east and west.

To explain this remarkable adjustment, *Guidebook 7*, as indicated in Figure 1, suggests that the Conococheague was flowing at or above the level of the Schooley peneplain in a meandering course about 1,500 feet above its present elevation. At that time it was meandering along the axis of the same syncline as now, but on weak Silurian shales between homoclinal ridges of the resistant Tuscarora sandstone. When the post-Schooley uplift occurred, the Creek is supposed to have cut downward through the ridge-making Tuscarora sandstone and the Juniata formation, respectively 270 and 400 feet thick,[4] and to have retained its inherited meandering course on the Martinsburg shale at the level of the Harrisburg peneplain. This explanation avoids the problem of the relative resistance of the limestone and shale. According to this solution, the Conococheague was formerly a subsequent stream flowing in a synclinal valley between parallel homoclinal ridges of the Tuscarora.

This interpretation seems unacceptable. While an earlier stream may have occupied the synclinal valley postulated, this valley would have narrowed in the course of regional reduction, and the two homoclinal ridges would tend to approach each other, finally to merge as a synclinal ridge. Thus the synclinal valley would have disappeared, and the stream would have been dismembered and entirely destroyed by headward-growing obsequent streams, according to well-recognized principles of folded mountain evolution. The vulnerable position of the earlier stream to such modification seems sufficient alone to rule out the hypothesis of an inherited course for the Conococheague.

In addition, it is to be doubted whether a stream cutting downward through 1,500 feet of rock between its former position on the Schooley peneplain and its present elevation would inherit so precisely adjusted a course on a lower formation. Furthermore, it is not likely that, if such a remarkable adjustment were attained by inheritance, it would have been maintained through two post-Harrisburg rejuvenations unless there were some special continuing virtue in a position on the shale.

To avoid these difficulties, the writer suggests that the Conococheague first achieved its present course as a new subsequent stream *during* the denudational interval *between* Schooley and Harrisburg time and that the Creek was already well adjusted to the shale by the time the Harrisburg peneplain had been made. This interpretation requires that the Valley limestones be weaker than

[3] The 1933 edition of the geological map of Maryland shows the meander bend to be about 0.3 miles west of the limestone-shale contact. The 1931 edition of the geological map of Pennsylvania, however, shows the contact at the state boundary, 1 mile north of the meander bend, about 0.5 miles farther west than does the Maryland map; while R. S. Bassler ("The Cambrian and Ordovician Deposits of Maryland," *Maryland Geol. Surv.* [1919], see map, Pl. I, and p. 156), who noted that the Conococheague was confined to the shale, depicts the bend and the contact as coinciding on his geological map.

[4] Stose, *op. cit.*, columnar section following p. 19.

the shales toward chemical attack but more resistant toward mechanical erosion by streams. Freeman Ward[5] has pointed out that the Somerville peneplain corresponds with remarkable exactitude to those areas underlain by limestone, while the higher Harrisburg remnants are, with similar exactitude, restricted to the shale. He has also shown that streams flowing from the shale to the limestone have broader floodplains in the shale than in the limestone, a fact pointing to the superior resistance of the limestone toward stream corrasion, if the formation of floodplains by lateral corrasion is accepted. These and other facts observed by Ward demonstrate well the opposite ways in which the shale and limestone of this region react to attack by chemical and mechanical agencies.

The most obvious explanation of the subsequent course of the Conococheague would seem to demand that during the Harrisburg cycle the climate be more arid than during the Somerville and that chemical reduction of the limestone be less effective than mechanical attack on the shale, so that a subsequent lowland was developed on the shale by differential denudation. With a more humid climate in Somerville time, the limestone areas were rapidly reduced by solution, while the shale areas remained largely at their former levels except where stream-dissected. This explanation seems adequate; but, with little supporting evidence pointing toward a sufficiently more arid climate in the eastern United States during Harrisburg time, it is difficult to accept. True, the Somerville peneplain is dated as beginning Pleistocene, a time generally taken as one of exceptionally high rainfall and atmospheric moisture. On the other hand, it cannot be definitely stated that the climate of the Harrisburg cycle was so much less humid that the limestone would then have acted as the stronger rock.

The present elevation of the Conococheague, at least 100 feet below the surface of the limestone country, suggests, however, that the Creek could have gained its subsequent course in a humid climate. It is evident that concentrated linear erosion by the Conococheague has been more effective in lowering it toward baselevel than have dispersed areal solution and stream erosion of the limestone. If this were not so, the limestone country, and not the Creek, would be the lowest feature in the landscape. Since in a humid climate the shale has thus been proved to be so susceptible to stream erosion that the Conococheague has maintained itself consistently at elevations below the limestone, it appears that streams developing on the shale have an advantage over those on the limestone. During the Harrisburg cycle the Conococheague was such a stream; and not only did it successfully extend itself along the belt of shale, but it entrenched itself so well that it became the master-stream of the shale and of the adjacent limestone country as well. Zones of limestone several miles wide were thus made tributary to the Conococheague by streams now rising in the limestone and flowing into the shale hills to reach the Creek, which is, for them, local baselevel.

While the Conococheague provides the most remarkable example of adjustment to the shale, it is far from being an isolated case. Conodoguinet Creek, rising north of Chambersburg, flows for the greater part of its course to the Susque-

[5] "The Role of Solution in Peneplanation," *Jour. Geol.*, Vol. XXXVIII (1930), pp. 262–70.

hanna River on shale, although the southern half of the Great Valley is here reduced to a limestone lowland. Swatara Creek, east of Harrisburg, flows for many miles along the shale belt, then, without the slightest deflection, crosses the limestone. In West Virginia and Virginia, Opequon Creek, flowing northeastward into the Potomac along the same shale belt as the Conococheague, shows an adjustment almost as perfect as the latter for about 30 miles along the strike. It is probably not without significance that the Potomac itself is offset about 6 miles to the southwest as it crosses the shale belt. Farther south in the Valley of Virginia, the same tendency of the streams to avoid the limestones is apparent, although it is less striking than in the instances cited.

The subsequent course of the Conococheague is of interest in itself and, when considered with other similar cases, strengthens Ward's conclusions that the Somerville peneplain was due to solution and that, in spite of its lower elevations, the limestone is more resistant than the shale toward mechanical erosion. Thus is explained the apparent anomaly in the position of many Great Valley streams.

MUNITIONS BEGIN UNDERGROUND

E. B. BAILEY
Director of the Geological Survey of Great Britain, London

At the outbreak of the war the Geological Survey, one of the many branches of Britain's Department of Scientific and Industrial Research, put aside all its usual preoccupations. It ceased to produce maps and memoirs, except for a restricted issue of wartime pamphlets created for the occasion. Instead, it reorganized its research and other activities so as to be in a position to give advice at short notice to appropriate authorities under three main headings, namely: (1) home mineral resources, to meet essential needs, further aggravated by the economy of shipping and foreign currency; (2) underground water, for new airfields, camps, and factories; and (3) subterranean facilities for storage and personnel.

One example was immediately afforded of the new value which accumulated knowledge may sometimes acquire under changed circumstances. London presented so obvious a target to the enemy that a call arose for prompt reinforcement of many of its buildings. The Geological Survey was asked to indicate interior sources where sand suitable for filling sandbags might be obtained. This it could do at once from its existing maps, pointing to particular parks and public spaces. The Civil Defence authorities have estimated that the resultant sandbagging of London cost £125,000 ($500,000) less than would have been the case if the sand had been brought from operating pits situated outside the built-up area. I quote this estimate merely to give an idea of the scale of the transaction. The essential needs of the moment were speed of acomplishment and noninterference with river, railways, and external roads required for military traffic or evacuation of children.

Until the war no sand worked in Britain was of the extreme purity requisite for the optical glass employed in many of the instruments used by the forces. This presented no immediate difficulty, for optical-glass sand was of very low peacetime value, since it could be conveniently shipped in bulk from Holland, Belgium, or northern France. In June, 1940, however, values changed abruptly —almost overnight.

Fortunately, the Geological Survey had, as long ago as 1921, located a home source of optical-glass sand of the highest quality. The occurrence is situated in a remote district, but its importance as a potential strategic reserve had been constantly stressed in the proper quarter. Thus, in the difficult days that followed the Battle of France, supply of optical-glass sand was maintained without any interruption.

Coal is at all times the main mineral asset of Britain. Its continuous extraction is forever bringing new facts to light. It is natural, therefore, in times of peace, that watching, recording, and advising on the progress of coal-field development occupy a major proportion of the Geological Survey's attention. For a period after the fall of France the output from established mines produced a superfluity; but, with armament factories coming daily into operation, this stage was soon

left behind, and any additional source had to be eagerly considered, if only it seemed likely to produce fuel at a low manpower cost.

Owing to the antiquity of coal-working in Britain, there were, at the outbreak of war, very few records of the extent to which near-surface coal had been extracted, for most modern collieries operate at considerable depth. A thorough investigation was therefore started; and it was soon found that an unexpected proportion of near-outcrop coal had been spared, especially in the Nottingham-Derby-Yorkshire coal field. Here, fortunately, the seams are inclined at so low and steady an angle as to make them ideal subjects for opencast working with modern mechanical excavators.

Opencast, or crop-working, has now been intensively developed. The Geological Survey and its sister-department, the Fuel Research Station, have given assistance—the geologists in location of sites, and the fuel researchers in estimation of quality. When a site has been suggested by the Geological Survey, it is explored by boring and trenching to prove the local conditions, including freedom from old working. If all goes well, production follows—production according to a carefully devised plan in which the surface is returned to agriculture as quickly and as effectively as possible.

Outcrop working started on a small scale toward the close of 1941. Up to the end of April, 1943, over 2,000,000 tons had been produced; by the end of December, this figure had risen to 5,500,000 tons.

In Britain iron ore comes second only to coal in importance. Iron ore today is mostly won from bedded deposits of the Jurassic Age in the English Midlands, exploited, as in the case of coal crops, with mechanical diggers.

The rocks of the Midland ironstone fields include a considerable proportion of clays and have often adjusted their disposition to meet changes of load introduced, very gradually, by the natural excavation, or erosion, of the local valley system. Much additional insight into these adjustments has recently been obtained, and the knowledge gained has helped quarry-managers in the laying-out of working faces. Other assistance has been given through identification of fossil bands in the rock cover. The outcrops of these bands, when mapped, furnish natural contour lines defining the depth at which iron ore is to be expected at any particular locality.

Low-phosphorous iron ores in Britain are almost all replacements or veins. The Geological Survey has guided successful development of certain previously abandoned veins in the Pennine metalliferous field.

Much attention has been given in a number of districts to veins of lead, zinc, tin, and wolfram, so as to assist in meeting war shortages. The spars which accompany the nonferrous metal area have proved of equal importance. The Geological Survey warned the steel industry of a possible fluor shortage and thus helped to avoid an industrial difficulty. One important step was to interpret the shape of a particular fluor body, which allowed of ordered replanning of mining operations.

Several other minerals could be listed which the Geological Survey has usefully investigated during the war including refractories for furnaces, feldspar for pottery, limestone or agriculture, and mica for electrical apparatus. In feldspar, Britain, through Geological Survey advice, has passed from zero production to wartime self-sufficiency, even in the

most exacting grades required for dentures.

The Survey has collected and correlated much geological information about wells sunk in Britain. Already during the war 11,500 entries for wells over 20 feet in depth have been made. The details given include the measured geological section and the diameter of each well and all available hydrological information regarding the depth at which water was struck, the rest-water level, the pumping-water level, the yield, and the quality. The Air Ministry has treated the Survey as its consultant in regard to sources of underground water for the many airfields now scattered over Britain.

REVIEWS

Structure and Evolution of Palestine with Comparative Notes on Neighbouring Countries. By LEO PICARD. (Bulletins of the Geological Department, Hebrew University, Vol. IV, Nos. 2, 3, and 4.) Jerusalem, 1943. Pp. 134; figs. 18.

The author explains in a Preface that this article is not in the form in which it was to have been presented before a Middle East Geological Congress, since the latter did not assemble, but is designed to serve two ends: namely, to inform fellow-geologists and also to describe the geology to the general public. The subjects differ, accordingly, not only in subject matter but also in treatment.

Part I deals with depressions and mountain structure. The relatively low Jordan Valley on the east and the coastal plains on the west are identified as longitudinal depressions. Two transverse depressions, the Emek on the north and the Beersheba on the south, transect the mountain mass, trending northwest. A number of minor depressions, "like big holes or scars, disfigure the otherwise uniform face of the landscape."

The depressions are said to be nearly all of tectonic origin. They are "depressed troughs, which generally originated in two considerable border faults and may be designated as Graben or rift valleys." They are surrounded on all sides by faults and form "radially downfaulted areas." Some other depressions are described as "fold-bends (monocline, syncline)." The distribution of these structures is summed up in the following statement:

> The mountains of northern Palestine, Samaria, and Galilee are rich in collapse depressions; whereas the south—the Judean highland from El Lubban to Beersheba and the Negeb—shows but little or nothing of the phenomenon, although there is no lack of faults in the southern area. The fractures and faults are younger structures, originating for the most part at the end of the Tertiary or the beginning of the Quaternary.

Under the heading "Mountain Structures" the author distinguishes folds, upwarps and downwarps, and fractures. His meaning is more explicitly set forth in the following, under the heading "Tectonics":

> We distinguish three fundamental tectonic forces: folding, warping, and fracturing. From these emanate the structural elements as follows:
>
> *Folding*, which leads to a series of anticlines and synclines, mainly of Middle Tertiary age.
>
> *Warping*, which causes later, i.e., in the transition from Tertiary to Quaternary, upwarps or arches and basins (anticlinorial and synclinorial regions).
>
> *Fracturing*, which produces double bordered rift valleys and horsts, or single bordered fault blocks and tilted blocks, mainly of Older Quaternary origin. There exist also blocks and depressions due to Tertiary faulting; but these faults are less pronounced morphologically, unless they are rejuvenated by Quaternary movements.

Following this summary of tectonic processes is an enumeration of the several structures pertaining to each type. There is an outline map of the structures and a series of cross sections. The map presents the well-known anticlines of the Palestine plateau, trending northeast-southwest. The sections express the author's interpretation of the faults as hading to the downthrow. He adheres throughout the paper consistently to the hypothesis of tension and collapse, in accordance with the views of Suess and Blankenhorn, and fails to mention the fact that the alternative explanation, which substitutes compression for tension, has been preferred by some other geologists who have studied the region (B. Willis, "The Dead Sea Problem," *Geol. Soc. Amer.*, Vol. XXXIX [1928] and "Welling's Observations on Dead Sea Structure," *ibid.*, Vol. XLIX [1938]). The reader might be left in ignorance of the fact that the older view had been challenged.

Part II treats of the "Evolution of Palestine," covering the paleogeography of western Arabia and adjoining lands from pre-Cambrian to Quaternary. The earliest recognized feature is a belt of sediments, which the author compares with the Molasse of Switzerland and interprets as the waste of a mountain range comparable to the Alps. It extended from Transjordania to Abyssinia in pre-Cambrian time. The heights having been reduced to a peneplain before the Paleozoic, the subsequent history is represented by advances and retreats of the sea across the western part of the continental mass. Block diagrams illustrate the author's concepts of suc-

cessive oscillations, back and forth, and the fluctuations of the shoreline from east to west and vice versa are indicated in a diagram for all periods since the Cambrian. It appears that the sea advanced across Transjordania into Saudi Arabia during the Ordovician, Gothlandian, and late Cretaceous and repeatedly invaded Transjordania. To the eastward of the shoreline at any time, extensive continental formations, chiefly of sands, were spread, and they now constitute the monotonous Nubian sandstone of various dates, including marine members.

The stratigraphy and paleogeography of post-Eocene Tertiary (Oligocene to Pliocene) are discussed in detail. Divergent views regarding the occurrence of lower Miocene, Burdigalian, in Palestine and Egypt are offered. Marine and continental Miocene formations are described and mapped. Similar treatment at greater length is given the Neogene and Pliocene, with excursions into Syria, Egypt, and Iraq.

A transition from Pliocene to Pleistocene is set off as the time when warping and "epeirogenic" movements resulted in general faulting from northern Egypt to Syria and produced the great grabens. The reader who wishes to follow a description of the details of faulting and folding will find them set forth in pages 81–84.

The Quaternary is presented in a discussion covering the last sixty-odd pages with a degree of detail which is out of proportion to the length of time involved but is justified by the importance of the local formations in relation to human history. Marine Quaternary, the development of the coastal plain, the nature of the calcareous sandstone, Kurkar—these are some of the more important items discussed. A useful table on page 102 summarizes the movements of the coastal plain and correlates them with anthropological epochs from Acheulean to Mousterian. The Quaternary of the Jordan Valley is accorded special notice, with a structural diagram. Climatic changes and Pluvial and inter-Pluvial periods are indicated. A chronological table of Pleistocene events sums up the record (pp. 122–23). Finally, a brief "Retrospect" condenses the long story from pre-Cambrian to the present in four pages.

In *Structure and Evolution of Palestine* Professor Picard has assembled the results of more than a score of years of exploration and painstaking observation. Students of the geology of the Middle East will regret that he has been obliged to present his findings in so condensed a form, but they will nevertheless find this work of notable value, especially in the subjects where he is most at home, stratigraphy and paleontology.

BAILEY WILLIS

"Surface Trace of the 1855 Earthquake." By M. ONGLEY. (In *Transactions of the Royal Society of New Zealand*, Vol. LXXIII [1943], pp. 84–89 [map and 12 photographs].)

"Wairarapa Earthquake of 24th June, 1942, together with Map Showing Surface Traces of Faults Recently Active." By M. ONGLEY. (In *New Zealand Journal of Science and Technology*, Vol. XXV B [1944], pp. 67–78 [2 maps and 15 photographs]).

Interest in fault scarplets and their distribution, with its bearing on earthquake frequency and long-range prediction in certain districts, has been growing for some years in New Zealand. This interest has been stimulated recently by the occurrence of an earthquake which shook the Wellington province, causing considerable damage, especially at Masterton, in the Wairarapa district, on June 24, 1942, and by the discovery that the cause of this disturbance may be traced to movement on a near-by fault, which shows surface displacement most strongly marked on hills about 3 miles north of Taueru and with 10 miles of the town of Masterton. The author describes the scarplet and associated features along this line of disturbance.

His investigations of the scarplets, formed when a large earth block was tilted contemporaneously with the occurrence of the great Wellington earthquake of 1855, were made some years ago and are now published with full references to early accounts of the events. The line traced by these breaks closely follows the base of the eastward-facing fault-scarp front of the Rimutaka Range from a point near Palliser Bay, continues northeastward through the Wairarapa district, parallel to the base line of the Tararua Range (these are western or backbone ranges of the North Island), and next swings farther to the east into the lower ranges east of the intermontane basin occupied by the Manawatu River and its tributaries. "Although the fault trace has not been followed from end to end, it has been followed so far and picked up at so many other points that there is little doubt that it is continuous." At its northeastern end it "runs into

the old fault" bounding the isolated Waewaepa Range on the eastern side, "where the beds have moved thousands of feet." In all but two of the many places where it has been observed the upthrow (in 1855) has been on the western side and has amounted to 3–10 feet. Commonly there is a scarp higher than this along the line of the fault, but it "can be seen to be composite, consisting of a fresher break about 6 feet and older breaks forming a grassed step 20–50 feet high."

Such facts indicate that the 1855 movement was the latest of two, or probably more, displacements on the same fault surface; but it is made clear that the majority of the scarplets along this trace were not on, or even very close to, the line of any of the faults of tectonic importance that bound major block features. The author does not discuss the question whether the newer line of break is one of gravity subsidence only rather remotely related to a thrust fault or whether there has been, on this line, renewal of upthrust along a surface parallel to, or branching from, an older major dislocation. He quotes contemporary accounts, however, of actual uplift (with tilting of a broad strip of country) which took place on the upthrow side of the new fault at its southern end. He describes a single exposure of the outcrop of a fault (a branch from the main fault?) of the older system as revealing "a low-angle thrust dipping less than 30°." This is at the south end of the area affected; and the fault is more or less parallel to, but not actually at, the base of the Rimutaka Range, being three-quarters of a mile east of it. The overthrust side consists of rocks similar to those in the range, "probably Palaeozoic, and the east side is light, loose river conglomerate probably Pleistocene."

The simple nature of the movement of 1855 and its immediate predecessors is most apparent where the resulting scarp crosses nearly level ground (see Fig. 2 [Ongley's Fig. 11]). There is little indication of scissors faulting along most parts of the scarp or fault trace, though Ongley notes that "in two places the east side has been raised 6 to 10 feet." At a number of the outcrops of the fault, however, a notch has been developed at the scarplet by fault movement with downthrow in the regular direction (to the east), because at these points the fault intersects spurs and foothills with westward surface slopes (see Ongley's Figs. 4 and 9). (There are some similar features along the lines of scarplets in the vicinity of Wellington city.)

In some of the numerous localities in which scarplets have been described, in both the North and South islands of New Zealand, to which the author refers in his Wairarapa paper, these are of the nature of "earthquake rents" or "reverse scarplets," as the reviewer now prefers to call them, which not only notch spurs but also develop trenches that extend for miles across spurs and ravines. These the author does not

FIG. 1.—Scarplet formed in 1942, 3 miles north of Taueru, Wairararapa district (*N.Z. Jour. Sci. and Tech.*, Vol. XXV, p. 70, Fig. 4). Photo by M. Ongley.

FIG. 2.—View looking south along scarplet 5 miles west of Masterton (rejuvenated in 1855) (*Trans. Roy. Soc. N.Z.*, Vol. LXXIII, Pl. 17, Fig. 11). Photo by M. Ongley.

distinguish from ordinary scarplets. Their nature indicates, however, that the latest movements on the major faults underlying them have been in a direction the reverse of earlier movements; and such features, as a whole, do not lend themselves to the alternative explanation that they are merely spurs notched in the manner described in the preceding paragraph.

In a map (7 × 12 inches), accompanied by an extensive list of references to authorities, which is issued with the Wairarapa paper, an

attempt has been made to show all known scarplets and earthquake rents indicative of "recent" earthquake-making fault movements in central New Zealand. These are all between Lat. 38° and 44° S. Quite justifiably, from the utilitarian point of view of long-range earthquake prediction, no distinction is made in the map between ordinary and reverse scarps. It may be noted, in passing, that the Hawke's Bay (1931) scarp is shown about 10 miles too far north: it passes south, not north, of the town of Napier.

The Masterton, or Taueru, scarplet (1942) crosses the top of a hill, peters out, and reappears beyond a small valley. It is very far from any ancient fault of tectonic or, at any rate, of topographic significance but has been the site of some small earlier (but recent) movement. Although fissuring and disturbance of the ground accompanied by slips and slumping of river banks close to the line of movement can be traced for some miles, well-marked scarps occur only along about three-quarters of a mile of an easterly to northeasterly trending trace. The new break makes a scarp which is nowhere more than about 3 feet high; but, since in places fresh cracks with this displacement are "in the foot of a face 10 feet high, it looks as if in the recent earthquake the ground has moved once more along a line where it had previously moved three or four times as much."

C. A. Cotton

David Dale Owen: Pioneer Geologist of the Middle West. By Walter Brookfield Henderson. ("Indiana Historical Collections," Vol. XXVII.) Indianapolis: Indiana Historical Bureau, 1943. Pp. xiii+180; figs. 7.

It is a pleasant change in these days of advanced, specialized research to turn to an intimate account of the first geological surveys undertaken in some of our middle-western states. Especially interesting this proves to be because a lively central figure—David Dale Owen—unifies several of these pioneer efforts in a connected story.

David Dale Owen, broadly trained in Scotland, Switzerland, London, and Cincinnati, was primarily a chemist until his thirtieth year. But in 1837 the Indiana General Assembly authorized an investigation of the mineral deposits of the state, and young Owen was selected for this. With great energy and enthusiasm he set out on a solo reconnaissance, differentiated the Carboniferous from the sub-Carboniferous (which at first included everything below the coal measures), came to suspect from the gentle dips that the coal beds found in Indiana were but a part of a large deposit that spread into Illinois in the form of a basin, and before the end of the year presented his report. This Indiana survey, however, soon gave way to a more urgent project.

For many years the policy of Congress had been not to sell but to lease public lands which contained mineral resources. The designation of specific mineral areas was in the hands of the surveyors of the Land Office, but, as none of them had the special training required for the work, mining reservations were located in a hit-or-miss fashion, many disputes arose, and trouble piled up. The problem became a major one after Black Hawk's defeat in 1832 when the land ceded by him to the United States was opened to settlement. Both miners and farmers rushed in, as northeastern Iowa and adjacent Wisconsin and Illinois were known to contain lead ore and were likewise desirable agricultural country. The demands of the two groups conflicted.

On August 17, 1839, Owen, at his home in New Harmony, Indiana, received notification of his appointment as principal agent to explore the mineral lands of the United States and, with it, detailed instructions for a complete survey of the region which might contain lead ore. This was all to be completed before winter set in. Working against time, Owen gathered together a score of his fellow-townsmen who had heard some of his popular lectures, proceeded to St. Louis, where several score additional men were hired, and then drilled his army on the fundamentals of geology and surveying on the steamboat trip up the Mississippi. They landed at Rockingham, Iowa, September 17, and promptly deployed in twenty-four squads. What followed is interesting reading. In spite of many difficulties, Owen's capable generalship brought the task to a successful completion on November 24, just before a severe snowstorm set in and the temperature dropped to 12 or 14 degrees below zero.

Probably the high point in Owen's career was the Wisconsin, Iowa, and Minnesota survey in 1847–49, which took him to places where no white man had been. He penetrated Canada, and an assistant reached the Bad Lands of South Dakota, bringing back a large number of fossils hitherto unknown. Naturally

only the main geologic features could be discovered in traversing such large areas, yet the results and their skilful presentation by the indefatigable Owen turned the eyes of American and British geologists toward various new questions. Congress had demurred over the high costs entailed by Owen's ambitious plans for printing, but in the end yielded to his insistence that his plates, sections, and maps required new standards of reproduction as well as quarto size instead of the octavo format used for government publications.

The last six years were the most crowded of Owen's overcrowded life. Along with various other activities he made a geologic survey of Kentucky in the first three and of Arkansas in the last three, in spite of failing health and recurring fever. As Indiana was then seeking him for a new survey of his own state, he drove hard to finish the Arkansas reports—too hard, in fact, for his life had been too busy, and he literally worked himself to death at the early age of fifty-three.

Owen was temperamentally, as well as by circumstances, a pioneer placed in a new country. His quick mind turned eagerly to new and unexplored fields, one after another of which he explored with amazing industry. Naturally, he made mistakes which a more careful man might not have made. But the latter would not have accomplished what Owen did. The mistaken ideas today are interesting in themselves. Actually he placed the nomenclature of western formations on a firm foundation and, in so doing, helped to standardize names throughout the United States. It is partly because Owen courageously insisted that our geologists renounce their individualism and accept the Cambrian and Silurian of Sedgwick and Murchison that today Americans and Europeans talk the same geological language because of a universal name system.

One closes this interesting and well-written book with a feeling of high admiration for the enthusiastic, many-sided David Dale Owen.

R. T. C.

Optical Crystallography. By ERNEST E. WAHLSTROM. New York: John Wiley & Sons, Inc., 1943. Pp. 206; figs. 209. $3.00.

This small book is divided into seventeen chapters. The first two serve as an introduction to crystallography; the next four deal with isotropics and elementary optics and the microscope; the following six with uniaxials; the succeeding four with biaxials; and the final chapter is a sort of summary laboratory outline. The point of view is that of the mineralogist and not the chemist. In general the treatment, which is nonmathematical in the main, is lucid and simple; the numerous line drawings, many of which are stippled, constitute the most valuable part of the work. Both indicatrix and ray-velocity models are employed. While, in detail, the book is quite different from Winchell's *Optical Mineralogy* (Part I), in general content and mode of coverage the two are so similar that it is surprising that one publisher should put out both of them.

Wahlstrom's book would be better without the first two chapters; the attempt to cover morphology in 8 pages is full of inaccuracies, in part because crystals are treated as if they were wooden blocks. Three projections are described in 6 pages; no later use is made of the gnomonic; the stereographic serves to illustrate a triclinic crystal (p. 130) but in out-of-date orientation; and the orthographic is employed for a too-brief treatment of skiodromes. Physical properties are covered even more succinctly in the 3 pages of chapter ii. The treatment of extinction angles is far too brief; description of the methods of determining sign from interference figures is unnecessarily extended. The subjects of conical refraction and dispersion of the optic axes are well handled.

In a number of places "the index of refraction of violet light" is mentioned. Light has a velocity, the substance being traversed has the index. "Quartz glass" (p. 34) should not be used to mean "silica glass." A substance does not have a higher index for red and a lower one for blue (p. 43); correct statements appear on pages 33 and 53. The quartz wedge does not resolve white light into its spectrum (p. 79). Some of the figures are incorrect; thus the upper figure on page 42 should show diverging rays extending to the left above the grain. In the left-hand figure on page 61 the rays leaving the crystal on the right should not show differing wave-lengths in air. The numbers are reversed on the third diagram, page 64. The rays on page 84 should show both smaller amplitudes and shorter wave-lengths in the crystal plate than are shown in air. On page 89 converging light striking the condenser emerges as parallel light. Such errors of detail are easily corrected. On the

whole, the book will appeal to many teachers. With the exceptions noted, it is a praiseworthy piece of work.

D. J. F.

"The Haplolepidae, a New Family of Late Carboniferous Bony Fishes: A Study in Taxonomy and Evolution." By T. STANLEY WESTOLL. (*Bulletin of the American Museum of Natural History*, Vol. LXXXIII [1944].) Pp. 1–122; text figs. 52; pls. 10.

The title of this study tends to obscure its actual scope. The report, by virtue of its analysis of the evolution of the actinopterygian fish, is, in reality, one of the most important papers on fossil vertebrates published during the last several years. In the first sections of the work the author describes in great detail a group of late Carboniferous fish which he refers to a new family, the Haplolepidae. Two genera—*Haplolepis*, with nine species, and *Pyritocephalus*, with five—are recognized as representatives of the family. Two familiar generic names of late Paleozoic fishes, *Eurylepis* and *Mecolepis*, both preoccupied, are shown to be synonymous with *Haplolepis*. Descriptions of the species are detailed; and the illustrations, which consist of reconstructions of the species, detailed sketches of actual materials, and reproductions of photographic plates, are excellent. Had the paper merely included the descriptive sections and illustrations, it could have been considered a sound contribution to the knowledge of Paleozoic fish. It is, however, in the application of the morphological data to studies of evolution that the work becomes of outstanding importance.

The most significant parts of the report are concerned with the evolution of the family Haplolepidae and with its bearing upon the course of development of the actinopterygians as a whole. It is shown that similar changes of skull and body appeared independently in different evolutionary lines not only within the family but also later among the "Holostei" and "Subholostei." These changes do not, however, appear in combination solely in one later family, which then might be thought to have descended from the Haplolepidae, but rather in various associations in several distinct groups. This raises the question of the validity of the customary use of such characters in classification.

The evolutionary concepts are elaborated in the section titled "Bearing on Holostean Evolution," in which the author concludes that the polyphyletic origin of the "Holostei" is clearly indicated. The "order" is not natural but has been derived from more than one Paleozoic family and represents an evolutionary grade of development reached independently by several groups. In short, Westoll regards *all* actinopterygians which are different from the palaeoniscids as "forming a fringe of evolutionary twigs and branches around a compact ancestral bush. . . . ." The unity of the Teleostei is such that this term is significant in classification, but neither the group "Holostei" or "Subholostei" can be considered valid.

A most interesting section is that designated "General Remarks on the Evolution of the Actinopterygians," in which the author analyzes the relationships of the fish to their environment on the basis of a study of the mechanics of swimming. The correlation of changes in adaptations with changing environment gives the basis for his concept that "parallelism" has resulted—in large part, at least, from similar environmental impact on various groups, particularly in the change from nonmarine to marine environments, rather than from comparable mutations of similar genes in related lines, as presented by Brough. Westoll's argument is strengthened by recognition of significant time intervals, implying "millions of generations," between the appearance of similar changes in different groups.

The final sections of the paper consider the biological and physical environment and the distribution of the family Haplolepidae. The haplolepids lives in nonmarine, shallow, somewhat stagnant, waters. The wide distribution of the family implies intercontinental connections between the areas in which its members have been found. The bearing of the distribution upon the theories of continental drift and the permanency of continents is considered briefly, with the author leaning, without decisive evidence from the haplolepids, toward the concept of continental drift.

The study has a concise, two-page summary, addenda of relevant work published since it was written, and an excellent list of references which are cited throughout the body of the text.

EVERETT C. OLSON

Mineral Resources of Minnesota. Edited by WILLIAM H. EMMONS and FRANK F. GROUT. (Minnesota Geological Survey Bulletin 30.) Minneapolis: University of Minnesota Press, 1943. Pp. viii+149; figs. 25.

Introduced by a brief discussion of the general geology of the state, this report gives concise accounts of the occurrence, origin, and utilization of Minnesota's various mineral resources. Of greatest importance and first in interest are the iron ores of the Mesabi, Vermilion, Gunflint, and Cuyuna ranges, which naturally receive the most attention. Already the cover of glacial drift stripped from the ore bodies of the Mesabi Range bulks about twice as large as all the material excavated in constructing the Panama Canal. By the end of 1941 this range had shipped 1,180,000,000 gross tons of iron ore, while the total for the whole Lake Superior region was 1,932,000,000 tons. About an equal quantity of proved high-grade ore still remains. Eventual use, by beneficiation, of the enormously greater masses of taconite, or cherty iron formation, is discussed in the light of possible competitive iron ores.

In Minnesota there is very little high-grade manganese ore (containing more than 45 per cent manganese), but the manganiferous iron ores of the Cuyuna Range may prove to be a large factor in the national manganese supply if and when foreign supplies are not available, although in normal times manganese cannot be produced from them profitably in competition with foreign ores.

The various other mineral resources are treated approximately in the order of their commercial importance.

R. T. C.

X-ray Crystallography. By M. J. Buerger. New York: John Wiley & Sons, 1942. Pp. xxii+531; figs. 252; tables 34. $6.50.

This work treats of the geometry of crystal structures and the detailed methods of its determination up to the level of finding the space group but not the atomic positions. Crystal physics and chemistry are not covered. Only those techniques are described which make use of single crystals and monochromatic radiation, with especial emphasis on moving-film methods. The necessary theory is developed in lucid fashion, using both vector algebra and more elementary methods. The work is an outstanding one, well-nigh indispensable to the beginner in this field.

D. J. F.

Japan: Its Resources and Industries. By Clayton D. Carus and Charles Longstreth McNichols. New York and London: Harper & Bros., 1944. Pp. xvii+252; photos 67; maps and charts 21. $3.50.

What were the resources and industrial status of Japan at the start of the present war, and what will be the role of Japan in the post-war world are questions of vital interest today. For answers the authors have presented the most authoritative information obtainable on the geographical, agricultural, animal, mineral, commercial, industrial, manufacturing, and human resources of the country, tracing expertly the development of modern Japan and portraying clearly its strong points, weaknesses, and prospects for the future. Much of this material they had been accumulating over a period of years; fortunately, they were able to gather and check considerable new information just prior to the war in spite of the evasiveness and deception of the Japanese.

As the fourth volume in the "Harper Geoscience Series," this is primarily a textbook, yet so much of its substance is of timely, human interest, attractively handled, that it is also a book for the studious general reader. The illustrations mostly are large and effective. Although in some the detail has not reproduced satisfactorily and sharper definition is to be desired, the effect on the whole is pleasing.

Geology is here involved only in the occurrence and production of coal, petroleum, metals, and economic minerals. For up-to-date information on these products and their future possibilities, given rather concisely, the chapter on mining may be commended.

R. T. C.

The Dedication of the State Natural Resources Building and the Illinois Mineral Industries Conference, Urbana, 1944. Illinois State Geological Survey Bull. 68. Pp. 305; figs. 92.

The splendid new State Natural Resources Building, housing the Illinois Geological Survey and the State Natural History Survey, was dedicated on November 15, 1940; but the volume containing the dedicatory addresses and the papers presented at the Illinois Mineral Industries Conference which was held in conjunction has only recently appeared. The principal dedication address, "Our Better Ordering and Preservation," was delivered by Dr. Isaiah Bowman.

Part II, forming the major portion of the volume, comprises the papers presented at the 1940 Illinois Mineral Industries Conference. These are grouped as follows: "Current Research on Coal," three papers; "Economics of Coal," four papers; "Symposium on Devonian Stratigraphy," thirteen papers; "Economics of Oil-Field Practice," four papers; and seven papers under the heading "Industrial Minerals." Although considerably less than half the speakers were members of the Illinois Geological Survey, the many-sided treatment of the state's mineral-resource problems reveals the broad scope of the Survey's investigations developed by Dr. Leighton.

Of outstanding value to the stratigrapher and paleontologist is the symposium on the Devonian stratigraphy of the central states, involving Illinois, Iowa, Missouri, Oklahoma, Arkansas, Tennessee, Kentucky, and Indiana—each treated by a selected authority. These are followed by discussions and a general correlation of Devonian formations in Illinois and the adjacent states by G. Arthur Cooper.

R. T. C.

Colorimetric Determination of Traces of Metals. By E. B. Sandell. New York: Interscience Publishers, 1944. Pp. xvi+487; figs. 73. $7.00.

The first 110 pages of this book, divided into four chapters, are given over to a general discussion of colorimetric trace analysis, covering methods, precautions, procedures, and reagents. Besides colorimetry and spectrophotometry, some use is made of fluorimetry. The remainder of the work treats of the detailed procedures used for determining traces of forty-five metals and the rare earths; each of these chapters has the twofold division of separations and methods of determination, and many of them also treat of applications. In over a dozen cases one of the "Applications" listed is silicate rocks; other headings under "Applications" include ores and alloys, soils, water, biological materials, etc.

The author has been quite successful in his aim "to present a limited number of methods which at the present time appear to be best suited for dealing with traces of metals." This book should be of great value to analysts working in this field, which is now of so much research interest.

D. J. F.

Soil: The Most Valuable Mineral Resource; Its Origin, Destruction, and Preservation. By W. H. Twenhofel. (State of Oregon Department of Geology and Mineral Industries, Bull. 26.) Portland, Oregon, 1944. Pp. 48; figs. 35. $0.45.

In his Eighth Annual Message to Congress, December 8, 1908, Theodore Roosevelt said: "When the soil is gone, men must go; the process does not take long." In this bulletin not only does Twenhofel express his belief that "erosion of the soils is the biggest problem confronting the farmers of the nation over a tremendous part of its agricultural lands," but he chooses to go further and regard it as "the biggest problem confronting the nation as a whole." While prepared primarily for the people of Oregon, this bulletin is designed to impress upon as wide a circle of readers as possible how seriously our most important resource is being depleted through ignorance and carelessness. The many striking photographs from various parts of the country should convince even the unimaginative reader.

The general problem is treated as basically one of geologic processes. To give the reader a correct understanding of the factors involved, the author discusses adequately, but as simply and clearly as possible, the formation of the mantle rock, rates of rock destruction in forming the mantle, and its erosion by wind and water. Free use of quantitative factual information makes the discussion concrete and practical. The last two chapters are "Erosion in the United States" and "Protection of Lands against Erosion."

The keynote of the work is well expressed in the summary: "The foregoing considerations show how geologic processes can be made to work for, instead of against, man; how the water can be made to conserve and not to destroy the soil. Thus, the geologic processes become the friends and not the enemies of man."

R. T. C.

VOLUME LIII

NUMBER 4

THE JOURNAL OF GEOLOGY

July 1945

DEVONIAN OF NEW MEXICO

FRANK V. STEVENSON
Guayaquil, Ecuador

ABSTRACT

Three Devonian formations have been recognized in New Mexico—the Canutillo, Sly Gap, and the Percha. New studies now make necessary a revision of, and several additions to, the Devonian system of New Mexico.

The sediments below the Sly Gap formation in the Sacramento and San Andres mountains, hitherto called the "Canutillo formation," from supposed correlation with the Canutillo formation in the Franklin Mountains of Texas, are here considered to comprise the Onate formation.

Time equivalency of the Sly Gap to a portion of the Devonian of Iowa is suggested by new paleontological evidence, the Sly Gap being younger than the Independence shale and slightly older than the Hackberry group.

"Contadero" is a new formational name proposed for a series of carbonaceous shales and gray limestones above the Sly Gap formation near the geographic center of the San Andres Mountains.

The Percha shale consists of two readily divisible members, although no disconformity exists between them. The name "Ready Pay member" is proposed for the "lower Percha," which is composed of black, fissile, nonfossiliferous shale and in most sections involves two-thirds of the total thickness of the formation. "Box member" is proposed for the "upper Percha," which is commonly composed of gray to green calcareous shales with intercalated lenses of limestone. The Box member is confined to the Mimbres Mountains, whereas the Ready Pay member is believed to occur throughout southern New Mexico and probably in the Franklin Mountains of Texas. The Contadero may be a facies of the Ready Pay member.

LOCATION OF THE AREA

Exposures of the Devonian in New Mexico are limited to four major mountain ranges in south-central and southwestern New Mexico (Fig. 1). These ranges are, from east to west, the Sacramento, San Andres, Caballos, and Mimbres (Black) mountains. The structure of the Mimbres Mountains is similar to that of the southern Rockies, whereas the other three ranges represent typical basin-range structures.

The Hueco and Franklin mountains of Texas continue the structural trend of the Sacramento and San Andres mountains of New Mexico. Mud Springs Mountain north of Hot Springs, although separated by the Rio Grande River from the Sierra Caballos, is part of the same structure. Cooks Peak, northeast of Deming, is the result of a laccolithic intrusion and is related to the Mimbres Mountains.

PREVIOUS WORK

The earliest work on the Devonian of New Mexico was confined to the southwestern quarter of the state, where there are numerous outcrops of the Percha Shale. This was done in connection with the extensive development of the ore-bearing Paleozoic sediments in the Mimbres Mountains (Black Range).

Although no formational name was given, the first specific reference to shales of Devonian age was made by C. H.

Gordon and L. C. Graton[1] in 1906. Gordon[2] in 1907 assigned the name "Percha shale" to the Devonian sediments in Kingston–Hillsboro–Lake Valley district (Fig. 1). No specific type section was designated, but the name was applied in general to the many exposures of the shale along the Rio Percha. Most workers prior to 1907 had combined the Devonian shale with the overlying Mississippian rocks.

The fauna of the Percha shale was described by E. M. Kindle[3] in 1909, much of it being a duplication of the fauna of the Ouray limestone of Colorado, described by G. H. Girty.[4]

The first reference to a fauna of Lime Creek age in New Mexico was made by C. R. Keyes,[5] who reported that the Lime Creek fauna occurs in southwestern New Mexico and southeastern Arizona. For the latter state Keyes undoubtedly had in mind the fauna of the Martin limestone, which had already been correlated with the Hackberry of Iowa by H. S. Williams,[6] a correlation

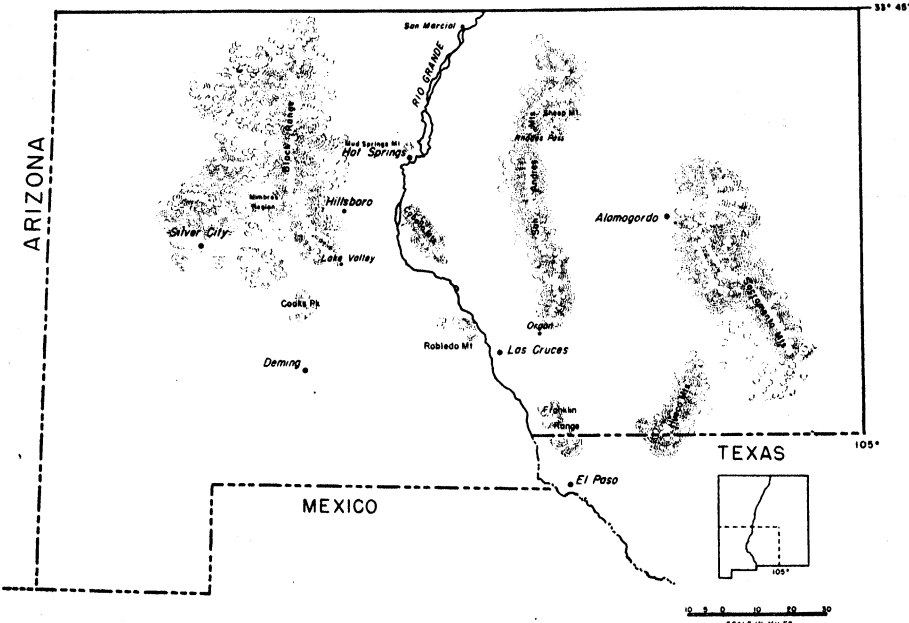


FIG. 1.—Index map of central and southwestern New Mexico

[1] "Lower Paleozoic Formations in New Mexico," *Amer. Jour. Sci.*, Vol. XXI (4th ser., 1906), pp. 394–95.

[2] "Mississippian (Lower Carboniferous) Formations in the Rio Grande Valley, New Mexico," *Amer. Jour. Sci.*, Vol. XXIV (4th ser., 1907), pp. 60, 62.

[3] "Devonian Fauna of the Ouray Limestone," *U.S. Geol. Surv. Bull. 391* (1909).

[4] "Devonian Fossils from Southwestern Colorado: The Fauna of the Ouray Limestone," *U.S. Geol. Surv. 20th Ann. Rept. (1898–1899)*, Part II, "General Geology and Paleontology," pp. 31–81.

[5] "Lime Creek Fauna of Iowa in Southwestern United States and Northern Mexican Region," *Proc. Iowa Acad. Sci.*, Vol. XIII (1906), pp. 197–9

[6] F. L. Ransome, "Geology and Ore Deposits of the Bisbee Quadrangle, Arizona," *U.S. Geol. Surv. Prof. Paper 21* (1904), pp. 38, 41.

supported in recent years by A. A. Stoyanow.[7]

Lists of the "Percha shale fauna" from the San Andres Mountains in southern central New Mexico were first made by N. H. Darton,[8] in 1917. Darton, assisted by E. Kirk, pointed out that "the beds in the San Andres Mountains appear to carry a smaller number of typical Ouray forms than are present at Lake Valley, and more forms characteristic of the Martin limestone and Nevada limestone."[9] These workers first recognized that the Devonian fauna in the San Andres Mountains and that of the Percha shale in the Mimbres Mountains are different in age.

M. A. Stainbrook[10] in 1935 suggested the possible existence of a formation or formations in New Mexico of a Devonian age other than the Percha. He presented several problems involved in the correlation of the Devonian strata in the Sacramento Mountains with the Devonian sediments in the San Andres and Sierra Caballos of New Mexico and the Franklin Mountains of Texas. In view of the inadequate field data at that time, Stainbrook's statement of the Devonian problems in New Mexico was remarkably complete. His paper also contained a plate illustrating fourteen genera of various species of fossils collected from a 30-foot vertical exposure of Devonian beds. On the basis of this collection he suggested a correlation with the Independence shale of Iowa.

The Sly Gap formation was named by the writer in 1941.[11] This formation included those sediments in the San Andres and Sacramento mountains previously discussed by Stainbrook.

Investigations made by the writer since 1936 led to the discovery of late Middle Devonian beds below the Sly Gap formation in the Sacramento and San Andres mountains.[12] This discovery prompted further investigations of the Canutillo formation, established by L. A. Nelson in 1937[13] and 1940,[14] in the Franklin Mountains of Texas. But in 1941 the writer did not correlate the Canutillo formation of Texas with the beds below the Sly Gap formation in New Mexico.

The following year G. A. Cooper suggested that the Sly Gap formation may include older Devonian beds of New Mexico.[15] This was a logical suggestion, inasmuch as Stainbrook[16] and L. R. Laudon and A. L. Bowsher[17] had reported the occurrence of *Leiorhynchus* sp. in the Devonian of New Mexico and since a similar *Leiorhynchus* sp. is found in the Canutillo of Texas. To expedite

[7] "Correlation of Arizona Paleozoic Formations," *Bull. Geol. Soc. Amer.*, Vol. XLVII (1936), pp. 486–87.

[8] "A Comparison of Paleozoic Sections in Southern New Mexico," *U.S. Geol. Surv. Prof. Paper 108-C* (1917), pp. 45–46.

[9] *Ibid.*, p. 46.

[10] "A Devonian Fauna from the Sacramento Mountains near Alamogordo, New Mexico," *Jour. Paleon.*, Vol. IX (1935), pp. 709–14.

[11] F. V. Stevenson, "The Devonian Sly Gap Formation of New Mexico," *Oil and Gas Jour.*, Vol. XXXIX (1941), No. 47, p. 65.

[12] Stevenson, "Devonian Formations in New Mexico," *Trans. Ill. State Acad. Sci.*, Vol. XXXIV (1941), No. 2, p. 163.

[13] Abstracts of theses for higher degrees, *Univ. Colo. Studies, Univ. Colo. Bull.*, Vol. XXXII (1937), No. 17, p. 89.

[14] "Paleozoic Stratigraphy of the Franklin Mountains, West Texas," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXIV (1940), p. 164.

[15] G. A. Cooper *et al.*, "Correlation of the Devonian Sedimentary Formations of North America," *Bull. Geol. Soc. Amer.*, Vol. LIII (1942), p. 1749.

[16] P. 711 of ftn. 10 (1935).

[17] "Mississippian Formations of Sacramento Mountains, New Mexico," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXV (1941), p. 2141.

description, the writer in 1942[18] correlated the Canutillo formation with the sediments below the Sly Gap in New Mexico. M. A. Fritz[19] conformed to this usage in 1944.

In 1935 K. C. Dunham[20] commented on two distinguishable members of the writer in 1941[21] and 1942[22] proposed "lower Percha shale" and "upper Percha shale" for the two members of the Percha shale.

Laudon and Bowsher[23] in their report on the Mississippian stratigraphy of the Sacramento Mountains in 1941 used the

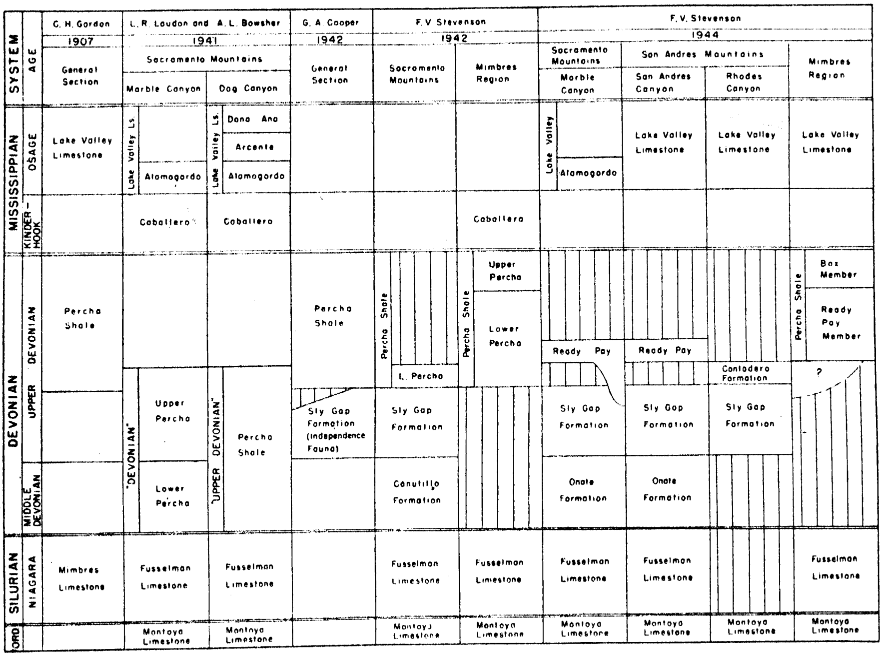


FIG. 2.—Correlation chart of the Devonian of New Mexico

Percha shale, but he did not elaborate on the division of the Percha shale or propose names for the two members. The term "Percha shale" for the Devonian formations in that area, with the subdivision into "upper and lower Percha shale" (Fig. 2). They recognized the complications arising from this usage, as well as the validity of the use of "Percha shale" in this area, as shown by the following quotation: "The recent work of

[18] "Devonian System: The Oil and Gas Resources of New Mexico," *N.M. School of Mines, State Bur. Min. and Min. Res. Bull. No. 18* (2d ed., 1942), pp. 23–24.

[19] "Upper Devonian Bryozoa from New Mexico," *Jour. Paleon.*, Vol. XVIII (1944), pp. 31–41.

[20] "Geology of the Organ Mountains; with an Account of the Geology and Mineral Resources of Dona Ana County, New Mexico," *N.M. School of Mines, State Bur. Min. and Min. Res. Bull. No. 11* (1935), p. 46.

[21] "Devonian Formations of New Mexico" (unpublished Master's thesis, University of Chicago, 1941), pp. 31–32.

[22] Pp. 23–24 of ftn. 18 (1942).

[23] Pp. 2107–60 of ftn. 17 (1941).

Stevenson in the area has demonstrated at least three distinct units in the so-called Percha formation, the upper of which is separated from the others by a distinct angular unconformity."[24]

THE CANUTILLO FORMATION

Varying opinion of stratigraphers regarding the correlation of the Devonian strata in New Mexico and Texas makes necessary the inclusion of a short discussion of the Canutillo formation.

The Devonian strata in the Franklin Mountains of Texas have been known for many years and were considered by Darton[25] as late Devonian ("Percha shale horizon") in age. Darton thought that the Devonian beds in the Franklin Mountains were correlatives of the Devonian sediments to the north, in the Sacramento and San Andres mountains of New Mexico. In a personal communication of Dr. G. H. Girty to Edwin Kirk, quoted in 1940 by L. A. Nelson,[26] the age was designated as "medial Devonian," although no formational name was given to the beds by Girty. Presumably the correlation of these Franklin Mountain beds with those in New Mexico was thus thrown open to suspicion.

L. A. Nelson[27] in 1937 had proposed the name "Canutillo formation" for the Devonian rocks in the Franklin Mountains and indicated their age as medial Devonian. He did not recognize this formation outside of the Franklin Mountains but in 1940 suspected that it may be found in the Hueco Mountains, Texas. He also stated[28] that the Canutillo formation had not been found in New Mexico. Nelson's description of the Canutillo formation is as follows:

> The Middle Devonian is represented in the Franklin Mountains by about 175 feet of sediments consisting of: cherty limestone, light brown color, immediately overlying the Fusselman; a thin bed of fossiliferous gray limestone; a thin bed of dense, almost black sandstone, which weathers brown; and about 40 feet of black fissile shale occurring at the top of the formation.[29]

The writer[30] in 1941 concurred with Nelson that the Canutillo formation does not occur in New Mexico. With the discovery of older Devonian sediments below the Sly Gap formation, however, and the occurrence therein of a species of *Leiorhynchus* similar to a *Leiorhynchus* sp. found in the Canutillo of the Franklin Mountains, and with the consideration of geographic proximity of these two formations, the beds below the Sly Gap in New Mexico were, in 1942 and 1943, correlated with the Canutillo formation.[31] This correlation met with the full approval of L. R. Laudon[32] but was rejected by L. A. Nelson.[33]

Unfortunately, little is known of the meager fauna of the Canutillo formation. It is presumably medial Devonian in age, but the faunal collections made by the writer are not sufficiently diagnostic to make any further statement at present. Study of the Canutillo strata at the type section indicates that they represent near-shore deposits, possibly from brackish water in part, since large plant remains have been found in the upper beds. It is possible that the Canutillo

[24] *Ibid.*, p. 2111.

[25] "Devonian Strata in Western Texas," *Bull. Geol. Soc. Amer.*, Vol. XL (1929), pp. 116–17.

[26] P. 164 of ftn. 14 (1940).

[27] P. 89 of ftn. 13 (1937).

[28] P. 164 of ftn. 14 (1940).

[29] *Ibid.*

[30] P. 163 of ftn. 12 (1941).

[31] Pp. 22–23 of ftn. 18 (1942); "Onondagan Equivalent in New Mexico," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXVII (1943), pp. 222–23.

[32] Personal communication.

[33] Personal communication.

formation is a time equivalent of the beds below the Sly Gap in New Mexico, even though they are not stratigraphic equivalents. Proof or disproof of this is difficult because of the lack of diagnostic faunas and the slightly different character of the sediments.

THE ONATE FORMATION[34]

"Onate" is a new formational name proposed for the beds below the Sly Gap formation in the Sacramento and San Andres mountains which were formerly identified as the "Canutillo formation." In honor of General Don Juan de Oñate, a late-sixteenth-century explorer and settler in this region, a ridge north of San Andres Peak was named "Onate Mountain." The term "Onate Mountain" was submitted to, and approved by, the Board on Geographic Names in 1944.[35] The type section (Fig. 3)[36] of the Onate formation is located on the north slope of San Andres Canyon, near an abandoned lead mine, in Sec. 18, T. 18 S., R. 4 E., San Andres Mountains, Dona Ana County, New Mexico (Fig. 4).

The Onate formation consists of a variable and intergradational series of shale, siltstone, fine sandstone, and limestone. Identification of these sediments for correlation purposes is complicated by lateral transitional changes, such as gradations from a shale to a siltstone to an arenaceous limestone in a distance of a few hundred feet. Additional difficulties arise from the lack of sufficient index fossils and the absence of strong color contrasts between the beds.

The base of the Onate is easily located, since the older Paleozoic formations, which are massive limestones and siliceous dolomites, are separated by an erosional disconformity from the relatively thin-bedded clastic beds of the superjacent Onate formation.

The top of the Onate is not marked by any disconformity readily detectable in the field. An outstanding difference between the Onate and the overlying Sly Gap formation, however, is the gray-brown color of the former and the tan to light-yellow color of the latter. This striking difference is due, in some degree, to a relatively high percentage of organic material in the Onate and to the occurrence of ferric iron and dolomite in the Sly Gap. Thin, shaly limestone beds, containing many ribbon-like bryozoa, described by M. A. Fritz[37] as *Sulcoretepora anomalotruncata*, occur within 3 or 4 feet of the top of the Onate. These bryozoa are found in most sections and form an excellent horizon marker for the near top of the formation.

The Onate has relatively less shale and more massive beds than the Sly Gap. In addition, flagstone bedding, accompanied by fucoids(?) and worm trails, are prevalent characteristics of the Onate but are rarely found in the Sly Gap. *Leiorhynchus* sp., occurring in great numbers in the Onate, has never been found in the Sly Gap.

The Onate formation is limited in outcrop to the Sacramento Mountains and the southern and central portions of the San Andres Mountains. Strata that may be related to the Onate are found near the northern end of the San Andres Mountains in the vicinity of Mocking-

[34] Pronounced: Own-yah'-tay.

[35] Personal communication. "The following decisions were rendered by this Board on April 13: Onate Mountain; New Mexico, Doña Ana County. A ridge in the San Andres Mountains north of San Andres Peak, between San Andres Creek and Andrecito Creek, T. 18 S., R. 4 E., New Mexico principal meridian. 32°45′ N., 106°34′ W."

[36] The writer introduces a symbol for siltstone in the stratigraphic charts, using a portion of the standard limestone symbol and sandstone.

[37] Pp. 32 and 40 of ftn. 19 (1944).

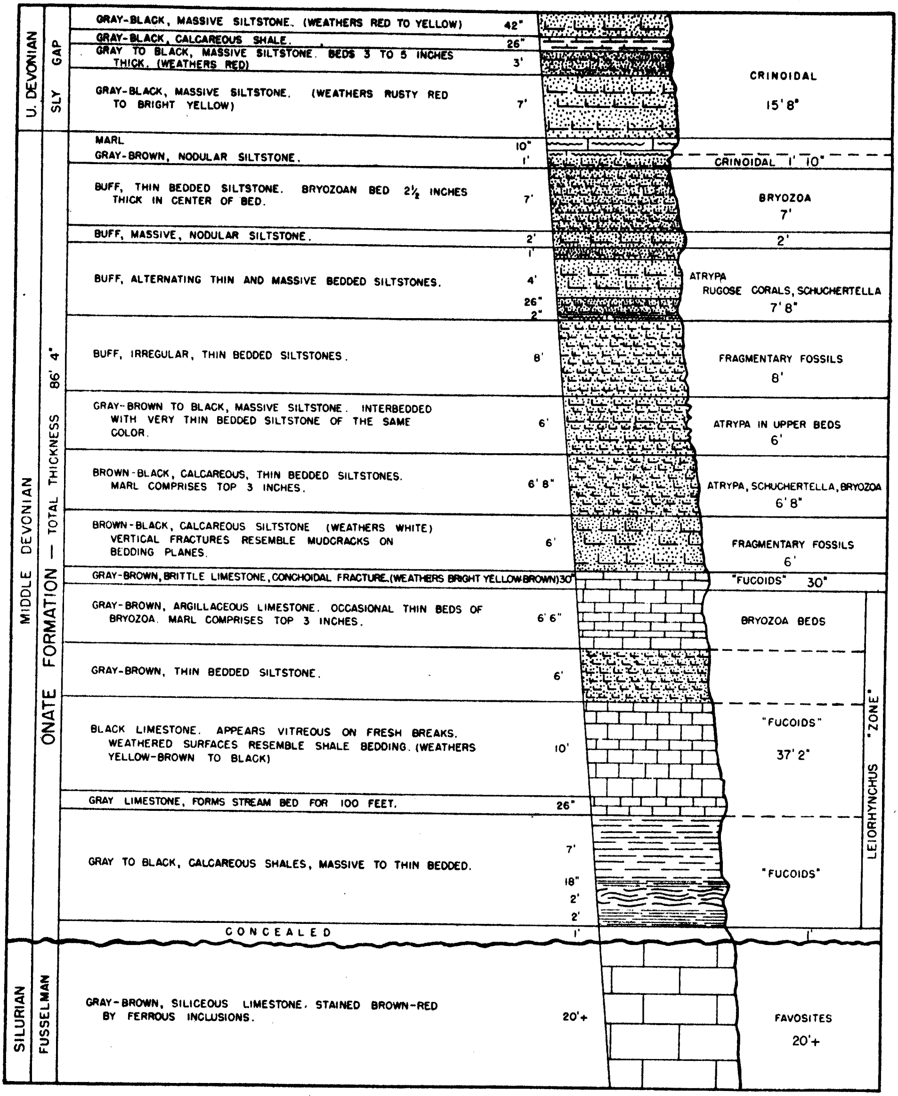
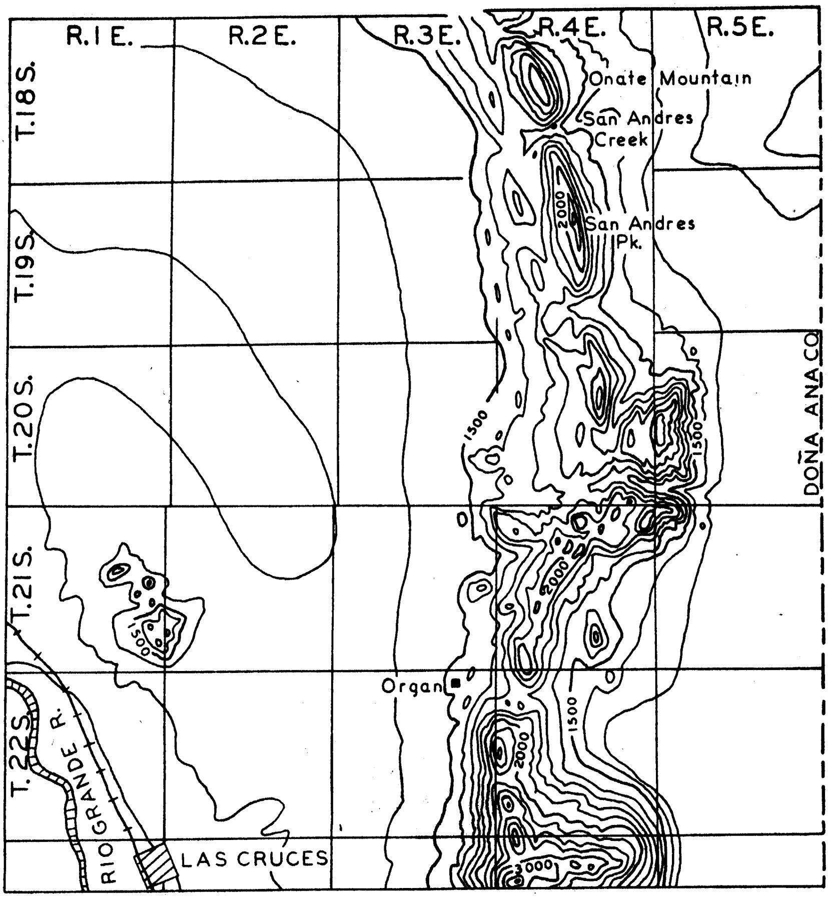


FIG. 3.—Type-section chart of the Onate formation



INDEX MAP SHOWING LOCATION FOR THE TYPE SECTION, ONATE FORMATION, SAN ANDRES CANYON, SAN ANDRES MTS.

SCALE 0 6 MILES
CONTOUR INTERVAL 100 METERS

BASED ON U.S.G.S. TOPOGRAPHIC MAP OF NEW MEXICO - 1925

FIG. 4

bird Gap.[38] Little is known of the character of these Devonian outcrops except that locally there are thin black cherts containing *Tentaculites* and *Paraspirifer acuminatus*, which indicate an Onondagan relationship. A description by Darton[39] of thin cherty limestones (Montoya formation—Ordovician) "overlain by a red shale of unknown age—possibly Percha" in the southern part of the Oscura Mountains, north of Mockingbird Gap, possibly refers, in part, to beds belonging to the Onate formation.

Topographically the Onate formation makes slopes and is covered by a veneer of talus. Where it is exposed on steep grades, the relatively thick beds weather along the bedding planes and have the appearance of man-made. steps (Fig. 5).

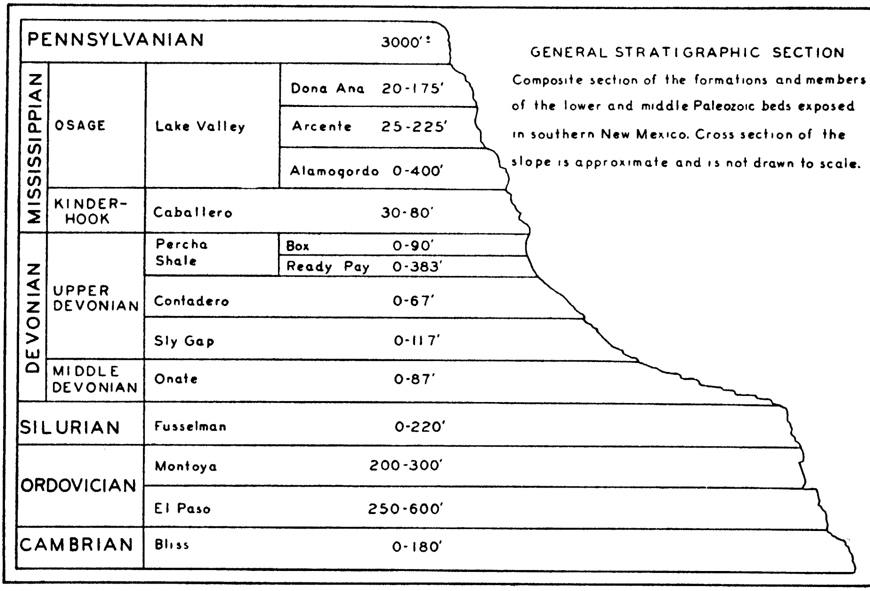


FIG. 5.—General stratigraphic section

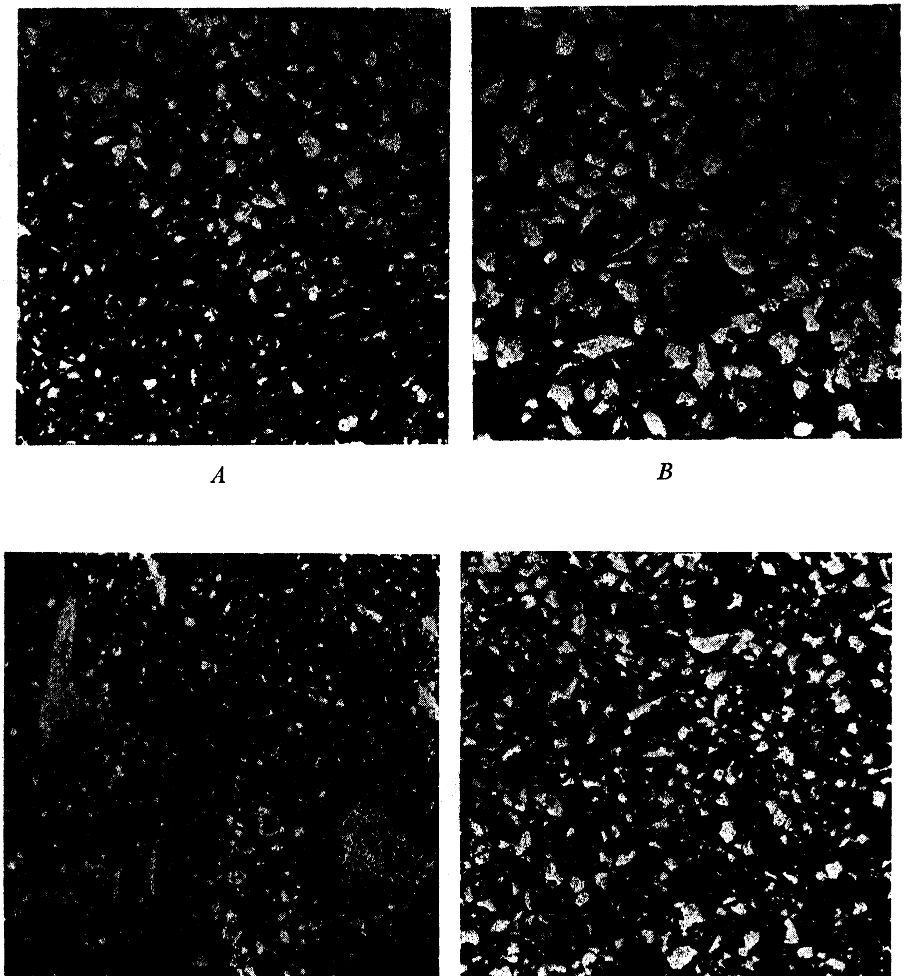
On the surface of the beds thus exposed, excellent specimens of *Leiorhynchus* sp. can be collected. The underlying formations—Fusselman in the Sacramento Mountains and the Fusselman and Montoya in the San Andres Mountains—are extremely resistant to weathering processes; as a result, the easily eroded Onate and the still less resistant Devonian formations above are commonly expressed in a benchlike indentation on most of the escarpments (Fig. 5).

The Onate type section (Fig. 3) in the San Andres Mountains exhibits a greater thickness than at any other known locality. The Onate thins to the north, although not so much as the underlying Fusselman limestone. In Rhodes Pass (Canyon) the Fusselman is missing; and the Onate beds, which are absent north of Rhodes Pass, lie unconformably on the Montoya limestone (Fig. 6).

The Onate formation can be traced ap-

[38] Stevenson, pp. 222–23 of ftn. 31 (1943).

[39] Darton, "'Red Beds' and Associated Formations in New Mexico," *Bull U.S. Geol. Surv.*, No. 794 (1928), p. 194.

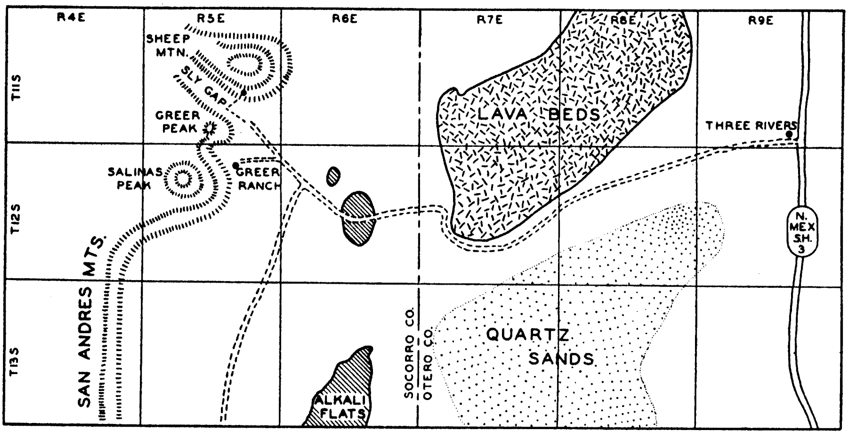


A *B* *C* *D*

Fig. 7—Microphotographs of Devonian rock: *A*, Onate; *B*, Sly Gap; *C*, Sly Gap; *D*, Sly Gap(?)

In southern New Mexico, ranch roads are annual and must constantly be kept passable. Therefore, localities established within the bounds of a large ranch will usually afford a certain degree of access. The locality also displays excellent weathered slopes for collecting, with a minimum of talus covering. In addition, there are also vertical sections for measurement and study. The fauna is more prolific than at any other locality investigated. Water is available for camping—an important point, inasmuch as it is necessary for anyone wishing to study the type section to expend at least two days in reaching, studying, and leaving the region. It is not advisable to visit this locality if there have been recent rains, as there are several playa areas which are dangerous to cross when wet.

The strata at the type section rest disconformably upon the dark-brown Montoya limestone, of late Ordovician age. Here a green porphyritic dike parallels the contact of the Montoya and the Sly Gap for approximately 100 feet. The Lake Valley limestone rests disconformably on the Sly Gap. The total thickness of the formation at the type section is 114 feet.



DETAILED MAP OF SLY GAP AND VICINITY

SHOWING A PORTION OF THE TULAROSA BASIN AND THE AREA IN THE IMMEDIATE VICINITY OF SLY GAP. TRAIL INDICATED IS ALTERED AFTER EACH RAINY SEASON, AND POSITION IS ONLY APPROXIMATE.

SCALE 0 — 6 MILES

BASED ON U.S.G.S., WATER SUPPLY PAPER MAP OF THE TULAROSA BASIN NO. 343

FIG. 8

The green porphyritic rock, indicated as *B* in Figure 9, is actually a dike, though at the type section it occupies the position of a sill. Many "greenstone" dikes intrude the Paleozoic sections in the southern region of New Mexico, and in many places such dikes parallel various formational contacts for several hundred yards before angling upward through the overlying beds. "Green-

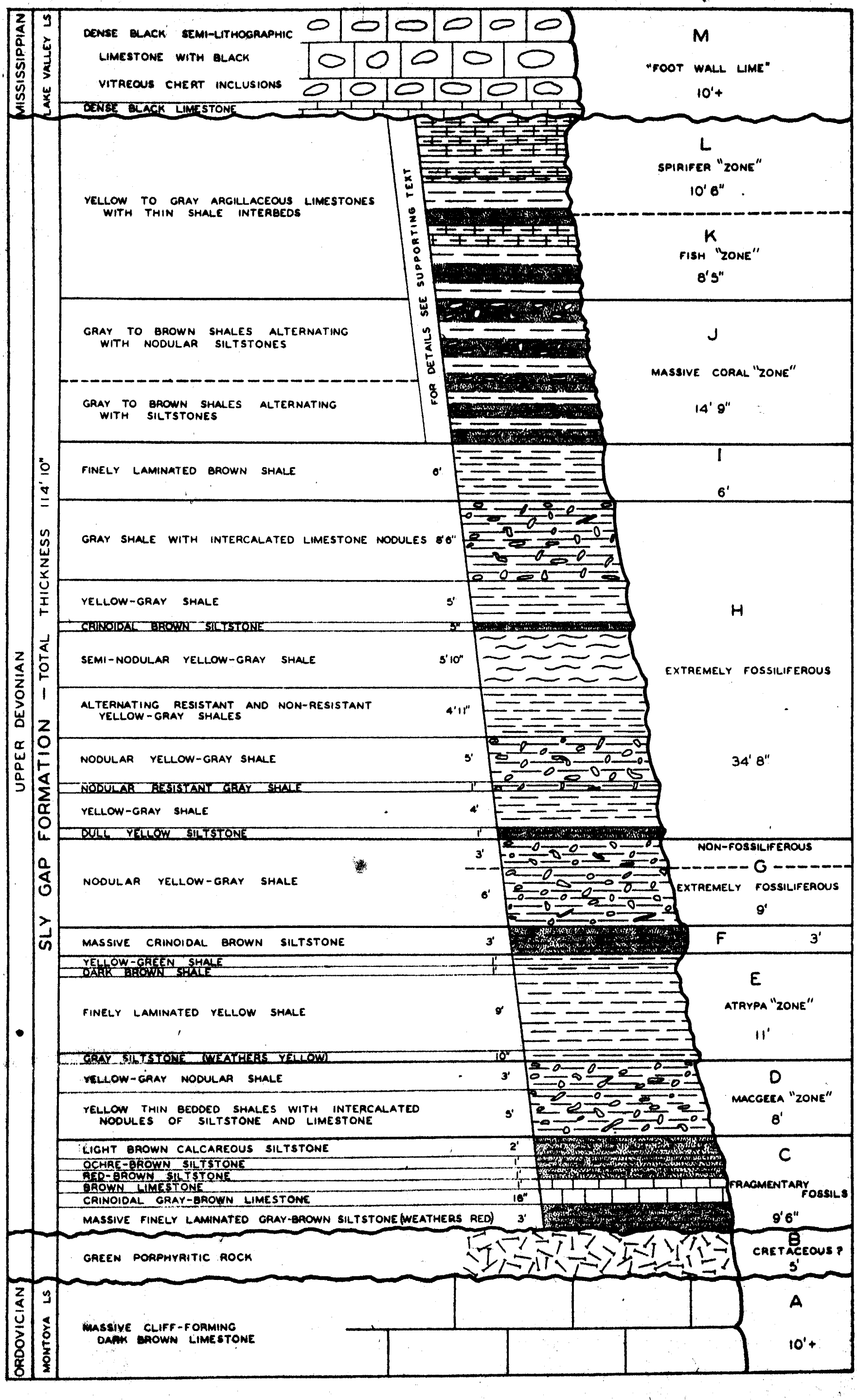


FIG. 9.—Type-section chart of the Sly Gap formation

stone" dikes in the San Andres, as well as in other mountains, have locally been found to mark both upper and lower limits of the Devonian as well as of the Mississippian.

The lithology of the Sly Gap formation makes adequate field description difficult. The formation is composed essentially of thin-bedded, alternating layers of shales, siltstones, and limestones. Rock samples, when tested in the laboratory, show a varying content of $CaMg(CO_3)_2$, $CaCO_3$, SiO_2, and organic material. Beds within the type section are not only very thin but unusually varied in lithology, thus necessitating the measurement of the section with a steel tape, inch by inch. An attempt has been made in Figure 9 to show the changing facies and the apparent effect on the fauna.

Precise "zone" designation by generic names is not entirely satisfactory because the faunules tend to intermingle. Thus, the faunal "zones" established indicate only an abundance of a single genus, which outnumbers all other forms in the same lithologic sequence of beds, and are not to be interpreted as true zones.

Zone *C*, as indicated in Figure 9, is everywhere represented in the Sly Gap formation and makes a red-brown marker for the Paleozoic section of the region where the Sly Gap formation is exposed. Crinoid columnals are commonly found in this horizon, although this characteristic is not so marked in the Sacramento Mountains as in the San Andres.

Wherever possible, there have been indicated on Figure 9, by names and letters, groups of beds carrying a predominance of certain genera. Beds essentially without fossils are indicated as "nonfossiliferous." Beds *G* and *H* are marked "extremely fossiliferous." These beds have many genera represented, without any genus being predominant.

Bed *J* is indicated as the "massive coral zone"; it contains *Alveolites*, *Hexagonaria*, and *Phillipsastrea*. At present this is the only known instance in which any part of the Sly Gap fauna is confined to a small part of the section. *Hexagonaria* and *Phillipsastrea* have been found only in the type section, but *Alveolites* occurs at other localities in the San Andres and Sacramento mountains. Colonial corals have not been found in the Sierra Caballos.

Bed *K*, indicated as the "fish zone," contains a few shark's teeth (see Sly Gap faunal list) and small, black, irregular-shaped nodules of undetermined origin. The strata are not very fossiliferous, and the forms other than those mentioned above are fragmentary.

Bed *L*, indicated as the "Spirifer zone," contains many poorly preserved specimens of *Cyrtospirifer whitneyi*. Fossils other than spirifers collected from this zone have been few. A 3-inch bed of calcareous shale at the top of the "Spirifer zone" marks the contact of the Sly Gap formation and the Lake Valley limestone. "Phosphatic" nodules are found in many places along this disconformable break between the two formations.

A thin section of a rock specimen from the "Spirifer zone" exhibits a field of 60 per cent quartz grains, averaging 0.07 mm. in the longest diameter; 5 per cent primary feldspar; and 35 per cent calcite, with occasional rhombs of dolomite (Fig. 7, *b*). The quartz grains are angular to subangular; well-rounded grains occur rarely. The grain size of the quartz approaches the lower limit of a very fine sandstone, and undoubtedly in places these beds should be called a calcareous sandstone. The hand specimen is sugary in texture; and it is friable, owing to poor cementation. The character of these sediments probably accounts for the poor

preservation of the fauna, as well as for its lack of variety.

For all general purposes, the outcrop area of the Sly Gap formation may be considered as confined to the Sacramento and San Andres mountains and the Sierra Caballo. The Rio Grande River marks the approximate western boundary of this area, and the easternmost exposures in the Sacramento Mountains give the east boundary. The north and south boundaries are within the limits of the aforementioned ranges. The region in which the Sly Gap formation is to be found extends, on the north, from longitude 105°55′ to longitude 107°20′ west; on the west, from latitude 32°30′ to latitude 33°30′ north.

In the San Andres Mountains the Sly Gap formation has been found exposed continuously from a point approximately 13 miles south of San Andres Canyon in the southern end of the San Andres to a point slightly north of Lava Gap in the northern end of the San Andres. The average thickness of the formation is 50 feet in the southern end of the range, and the maximum of 117 feet is found near Sly Gap (Fig. 6).

The Sly Gap formation overlies two different formations in the San Andres Mountains: the Onate, in the southern and central portion, and the Montoya, north of Rhodes Pass (Fig. 6). Two formations are found above the Sly Gap in the San Andres: the Ready Pay member[42] of the Percha shale in the southern and central portion of the range, except where the Contadero formation[43] apparently replaces the Ready Pay; and, at the extreme north end of the San Andres in the vicinity of Sly Gap, the Lake Valley formation, which is found above the Sly Gap formation.

[42] See p. 241.

[43] See p. 239.

Undoubtedly, the best exposures of the Sly Gap are found in the San Andres Mountains. Every canyon and gap (fault-formed) displays clearly the stratigraphic relations and provides fine collecting. The absence of roads and of maps provides the two main handicaps for work in this area.

The Sly Gap formation ordinarily forms an indentation or bench in the lower Paleozoics, similar to the Onate; and in most localities the strata are covered with debris from the overlying Mississippian and Pennsylvanian beds. Where veneering talus is not present, the lighter dun color of the Sly Gap rocks contrasts against the darker beds above and below. Aerial photographs are a great aid in locating new exposures of the Sly Gap because of this contrast in color. They were not used in the Sacramento Mountains but were employed in the northern part of the San Andres and in regions to the west.

SACRAMENTO MOUNTAINS

Exposures of the Sly Gap strata in the Sacramento Mountains display the same general characteristics as those in the San Andres, but it is notable that the sections are thinner in the Sacramentos (averaging 45 feet) and that there is a gradual thinning of the formation to the east and south in this range. Whether post-Devonian erosion or less deposition in Sly Gap time accounts for this thinning is not at present known.

The Sly Gap sediments are well exposed in the Sacramentos; and, with the possible exception of exposures in Rhodes Pass, San Andres Mountains, they are more readily accessible in the Sacramentos. Canyon sections investigated in the Sacramentos are, from north to south, as follows: Marble, Alamo (Fig.

10), Caballero, Gordon, San Andreas,[44] Dog, Escondida, Agua Chiquita, and Nigger Ed canyons.[45]

On the north wall of the first unnamed canyon north of Marble Canyon the Sly Gap formation crops out on the west flank of a northward-plunging anticline whose strike is N. 10° W. This anticlinal structure flattens north of this point, and only Pennsylvanian and younger beds are found beyond it. The Devonian sediments do not crop out south of Nigger Ed Canyon, near the southernmost extension of the mountains. The structural trend of the Sacramentos is known to continue south in the Hueco Mountains of Texas, but this area has not been investigated by the writer.

Faunal and lithologic zones of the Sly Gap formation are not so consistent throughout the Sacramentos as they are in the San Andres, and with thinning of the formation the stratigraphic units present a somewhat different aspect. More black-shale beds are incorporated in the Sly Gap of the Sacramentos than are found in the San Andres. The probable significance of this feature will be pointed out in the conclusions.

Faunal assemblages collected from various canyons more than a mile east of the western escarpment of the Sacramento Mountains do not display the abundance or the variation exhibited in

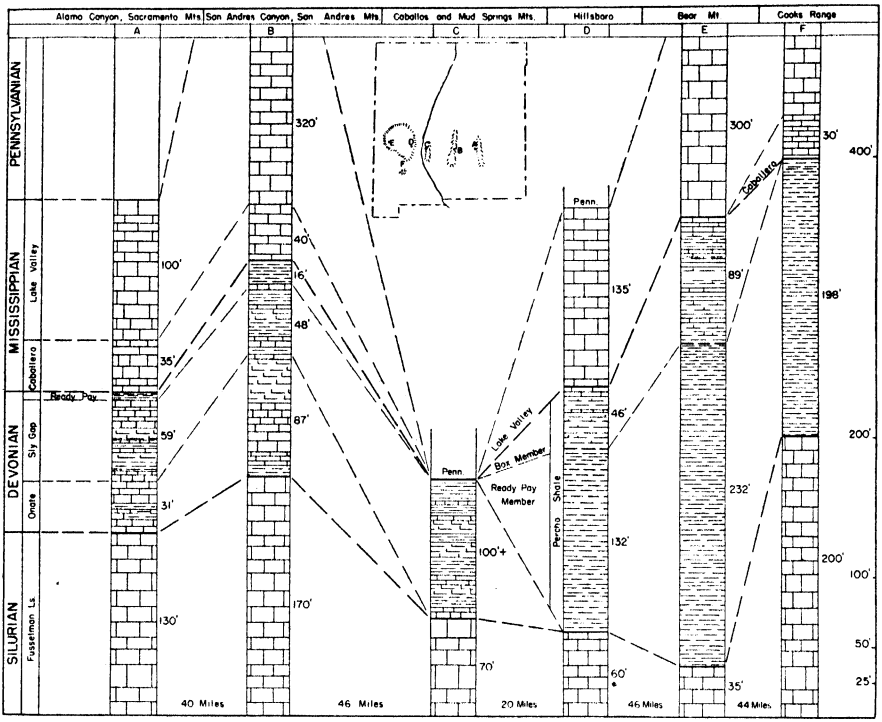


FIG. 10.—East-west correlation of the Devonian in southern New Mexico

[44] Not to be confused with San Andres Canyon, San Andres Mountains.

[45] I. S., Division of Grazing, and Forest Service surveys, 1940.

the collections taken nearer the escarpment itself. The incomplete collection of Devonian fossils figured in Stainbrook's report[46] on the Devonian of the Sacramentos is from the Sly Gap formation. This collection is valid with the possible exception of a figured specimen of "*Microcyclus?* sp.," which has not been found by the writer in the Sly Gap formation. Similar specimens have, however, been collected from the Caballero formation and identified by L. R. Laudon and A. L. Bowsher[47] as "*Hadrophyllum romingeri.*"

The Devonian section exposed in Dog Canyon presents several puzzling aspects. Locally the Devonian strata are massive and composed of dark-brown, calcareous siltstone approximately 45 feet thick. There is a meager fauna consisting of unidentified fragments and of species of *Schizophoria* and *Leiorhynchus*, which, on the whole, resembles the fauna of the Onate formation more than that of the Sly Gap.

A thin section from the upper beds of the Sly Gap formation in Marble Canyon (Fig. 7, *c*) shows a fine-grained, extremely fossiliferous limestone, with less than 10 per cent quartz grains and rare dolomite rhombs. The strata of the "Spirifer zone" apparently are not represented in the Sacramentos, and the upper strata exposed in Marble Canyon are probably equivalent to the upper beds of Zone *H* of the type section.

A structure of undetermined origin which reveals the only angular unconformity known between the Sly Gap and the overlying Lake Valley formation has been found in Marble Canyon (Fig. 11), which has two main branches—one draining from the northeast and one from the southeast. An elevation, V-shaped in plan with apex pointing almost due west, lies between these two tributaries. The structural feature is shown on the south slope of this elevation in an exposure approximately $\frac{3}{4}$ mile east of the quarry in Marble Canyon. The angular unconformity is found on the west flank of the plunging anticline noted near the beginning of this section. At this place the dip of the Sly Gap beds is 16°, Figure 11 being a view looking northward essentially along the strike.

Out of a total of 48 feet of Sly Gap sediments measured, approximately 45 feet are shown in the area labeled *A* in Figure 11. Not shown in the photograph are 32 feet of the Onate formation, below which the Fusselman limestone lies disconformably. The dip of these older formations is also 16°.

The area marked *B* in Figure 11 shows the Lake Valley strata with a dip of 13° in angular contact with the Sly Gap formation below. The 3° greater dip of the Sly Gap causes a few feet of its topmost beds to pinch out east of the area photographed. The Caballero formation, ordinarily found above the Sly Gap in the Sacramentos, does not crop out in this locality.

In area *C* there are at least 50 feet of black fissile shale, with a few thin stringers of arenaceous shale stained red-brown by oxidation of pyrite. The dip of the black shale varies from a minimum of 28°, near its contact with the overlying Lake Valley, to almost 90°, in its lateral contact with the Onate and Fusselman formations farther down the slope. The strike of the black shale is N. 10° E.

Talus and vegetation obscure the contact with the Fusselman farther down the slope, on the left of the observer. The angular contact of the shale with the Fusselman is probably lost in a relatively short distance, inasmuch as another out-

[46] Stainbrook, p. 712 of ftn. 10 (1935).

[47] Laudon and Bowsher, p. 2124 of ftn. 17 (1941).

crop, perhaps 200 feet to the southwest, exhibits a normal succession of beds without indication of angular unconformity. Except for this very local area,[48] as previously reported, the relation of one Paleozoic formation to another formable with the overlying Lake Valley, it is clear that the shale is post–Sly Gap and pre–Lake Valley.

The black shale shows no internal slickensides and no slickensided contact with the Sly Gap, Onate, or Fusselman.

FIG. 11.—Devonian structure in Marble Canyon, Sacramento Mountains

throughout the state of New Mexico is one of parallel disconformity.

Within certain limits the age of all the beds involved in the structure can readily be determined. Since no black shale older than the Sly Gap is known in this area and since the black shale is noncon-

[48] The structure under consideration fails to appear in the cliffs on the north side of the narrow V-shaped (in plan) elevation.

The individual shale beds maintain a uniform thickness regardless of changes in dip or proximity to the adjacent formations. The overlying Lake Valley limestone has not been deformed except for local sagging over the shale. It thus can be definitely stated that the structure is pre-Mississippian. The slight angularity of the Sly Gap and older formations with the overlying Lake Valley

and the eastward tapering-out of the upper beds of the Sly Gap are proof of an erosional period following the tilting of the pre-Mississippian formations. Because the same erosion that beveled the Sly Gap also beveled the black shale, there must have been a relatively thick black shale deposited above the Sly Gap prior to Mississippian deposition. In addition to the 50 feet or more of black shale preserved in this structure, beds of somewhat similar, black, fissile shale, ranging from a few inches to several feet in thickness, have been found above the Sly Gap and below the Mississippian in the Sacramento and San Andres mountains.

Several explanations for the structural relations have been considered, but none is entirely satisfactory. The writer has visited the structure on four occasions, accompanied by L. R. Laudon, J Harlen Bretz, and A. L. Bowsher. The latter participated in its discovery in the summer of 1939. Several others have also visited the area but have been unable to reach a solution to the problem.

One of the first theories advanced was that the black shale records a channel-filling. However, it is impossible to regard the 90° dip of the shale[49] as the result of any known type of initial deposition. Another factor to be considered is the localized tilting of the Sly Gap and older formations. Since this condition is not found at other localities, it would be something of a coincidence if channeling and tilting were to have occurred only once and at the same location.

There is also the possibility that the black shale was deposited against a steep post–Sly Gap erosional surface and that subsequent slumping of the black shale on this surface produced the increased dip. There is no evidence, however, to support this hypothesis, for reasons pointed out above.

It is possible that a cavern had developed in the underlying limestones and that, in a collapse of the roof, the black shale slumped into the cavity. When the very localized area of the structure is considered, this may be the explanation. But it must be pointed out again that the exposed contact of the black shale with subjacent older Paleozoic beds is nearly as straight as if laid out with a tape. Jointing of the limestone is rare, and no appreciable solutional enlargement of such joints as occur in the Sacramentos has been noted. Nor are sinks or caverns known anywhere in the formations under consideration.

Another possible theory invokes submarine landsliding, in which at least all formations from the black shale down through the Fusselman were involved. This is possible but not probable, for such movement should have distorted the shale layers, which would not have retained their uniformity of thickness. Contamination of the shale with fragments of other rock would also be expected. Furthermore, there is little evidence that the Devonian sea bottom was sufficiently irregular to afford the gradient necessary for so much vertical displacement as is indicated in Marble Canyon.

The writer believes that the structure in Marble Canyon was formed through some other process, which permitted the black shale to slump or drop. Normal or high-angle faulting seems most nearly adequate. However, this interpretation demands a fault whose maximum surface trace can hardly exceed 350 feet, while its vertical displacement is at least 140 feet. In the last four years the exposure has been considerably enlarged

[49] The part of the section showing this steep dip did not become exposed until 1942.

by erosion, and it may well be that in the future the talus covering will be removed and further data can be gathered.

MUD SPRINGS MOUNTAIN AND THE SIERRA CABALLOS

The finest exposure in the westernmost extension of the Sly Gap formation is found at Mud Springs Mountain, 2 miles northwest of Hot Springs, New Mexico (Fig. 1), although the section is rather difficult to locate, owing to the folded and faulted complex of the region. The Sly Gap strata here total 48 feet in thickness. They rest disconformably upon the Fusselman limestone and are overlain disconformably by the shale and cherty limestone of the Derry series of the Pennsylvanian system.[50] Much of the Upper Devonian in this area has undoubtedly been removed by pre-Pennsylvanian erosion, as evidenced by abrupt variations in the thickness of the upper shales of the Sly Gap.

The Sly Gap formation is also exposed on the westward-facing escarpment of the Sierra Caballos, south of Hot Springs. The Sly Gap beds found in the Caballos are similar in general character, thickness, and disconformable relations to those exposed in Mud Springs Mountain.

The fauna is the same as that found in the Sly Gap to the east, with the exception of a fossil tentatively identified as *Receptaculites* n. sp., which occurs in large numbers in the Mud Springs locality. The individual fossils are also much larger than those found to the east, but undoubtedly they belong to the same species.

The Mud Springs section is geographically nearer the Percha shale outcrops than any other exposure of the Sly Gap. There is no evidence, however, of any Percha shale or fauna being present. But there is an increase in the amount of shale toward the top of the section, and possibly it can be presumed that the Percha shales were once deposited over this area. The Mississippian limestones that ordinarily have protected the Devonian from excessive erosion were not deposited in this region or were removed by erosion prior to Pennsylvanian deposition.

Due west of the Sierra Caballos, in the vicinity of Hillsboro, New Mexico, are located the easternmost outcrops of the typical Percha shale. Paleozoic formations are not exposed in the Rio Grande River drainage area between the Sierra Caballos on the east and the Hillsboro region to the west; therein lies the difficulty in establishing the relationship of the Sly Gap to the Percha shale. Investigations of the Paleozoic section exposed in the Robledo (Roblero) Mountain, northwest of Las Cruces, did not reveal any Devonian sediments.

PALEONTOLOGY OF THE SLY GAP FORMATION

The only noteworthy studies of the Sly Gap fauna have been made by Stainbrook[51] and Darton.[52] Stainbrook's figured fauna from Gordon Canyon, Sacramento Mountains, includes only fossils found in the Sly Gap assemblage, with the exception of the previously noted "*Microcyclus?*" Darton's list of Devonian fossils from the San Andres Mountains undoubtedly included the fauna of the Onate as well as that of the Sly Gap.

Stainbrook[53] listed, but did not figure, the genera "*Chonetes*, *Ambocoelia*, *Lingula*, *Crania*, and *Liorhynchus* (sic.)" from "Bed 6" in the stratigraphic column,

[50] M. L. Thompson, "Pennsylvanian System in New Mexico," *Bull. N.M. School of Mines*, No. 17 (1942), p. 37.

[51] Pp. 709–14 of ftn. 10 (1935).

[52] Pp. 44–46 of ftn. 8 (1917).

[53] Pp. 710–11 of ftn. 10 (1935).

with which, however, his supporting text fails to conform. Furthermore, the genera listed are not found in the writer's Sly Gap collection or in the collections borrowed. *Ambocoelia* sp. is found in large numbers in the Contadero formation above the Sly Gap in the San Andres Mountains, and it is possible that the specimens of this genus came from Contadero strata. *Leiorhynchus* sp. is found in the Onate formation but has not been found stratigraphically higher. Stainbrook's "Bed 6" occurs beneath the Mississippian and above beds carrying the Sly Gap fauna.

The faunal list on page 239 includes only those genera and species found at Sly Gap itself. No attempt is made to list the complete fauna of this formation, for investigation is still in progress, and a much more detailed descriptive paper will be published as a separate unit. It is significant, however, that the Sly Gap locality does yield approximately 95 per cent of the total fauna, and many forms occurring there are missing at other localities. The faunal richness of the type section may be judged by the fact that, in approximately five hours of collecting, the writer and A. L. Bowsher collected over five thousand specimens from it.

The most diagnostic and consistently present fossils found in the Sly Gap formation are representatives of the genera *Atrypa*, *Productella*, *Nervostrophia* (*Leptostrophia*), *Emanuella?* and *Macgeea*. These five genera are always found in association in every well-exposed section of the formation, and their species occur in larger numbers than do those of other genera.

The genus *Emanuella?* has a cyrtinoid shape and is probably "related to the general group of *Martinia* or *Emanuella*,"[54] with the possibility that the individuals referred to it represent a new genus. This form was figured and identified by Stainbrook[55] as "*Ambothyris?*"

The genus *Macgeea* has been thin-sectioned and is definitely identified as *M. solitaria*, which has been found in the Hackberry of Iowa. From the exterior, *M. solitaria* resembles *M. parva*, which occurs in the Independence shale of Iowa. Stainbrook[56] identified and figured a similar, if not identical, coral as *M. parva*.

Atrypa is the most abundant genus in the Sly Gap fauna and is represented by at least four species: *A. rockfordensis*, *A. devoniana*, and probably two new species.

Both in number and variety the Sly Gap fauna consists predominantly of brachiopods. The corals make up less than 5 per cent of the fossils collected, although *M. solitaria* is one of the five most abundant fossils. One brachiopod specimen from the Sly Gap is identical with *Stropheodonta inflexa*, found in the Snyder Creek formation of Missouri. Other genera, not specifically identified, indicate a close specific affinity to many fossils found in the Hackberry and the Independence shale of Iowa.

Colonial corals from the Sly Gap have been sectioned and are conspecific with *Alveolites rockfordensis* from the Hackberry. Other corals are identified as belonging to the genera *Phillipsastrea* and *Hexagonaria*. Colonial corals have not been reported from the Independence, but they do form an important part of the faunal assemblages of both the Sly Gap and the Hackberry.

The crinoid genus *Arthroacantha* has been found in both the Sly Gap and the Independence shale. The Independence

[54] G. A. Cooper, personal communication.

[55] Pl. 83, Figs. 16–18, of ftn. 10 (1935).

[56] *Ibid.*, Pl. 83, Figs. 8, 25–28.

form was described by A. O. Thomas[57] as *A. mamelonifera;* but the Sly Gap species, though closely related to the Independence form, is new. The occurrence of this rare late Devonian crinoid in the Sly Gap indicates a possible relation to the Independence fauna.

The cephalopod *Manticoceras regulare* has been found in both the Sly Gap and the Hackberry.[58]

The writer believes that the Sly Gap fauna is slightly older than the Hackberry and younger than the Independence fauna. The paleontological evidence also indicates that the fauna of the Sly Gap is slightly older than that of the Martin limestone of Arizona, although many of the genera represented in the two formations are the same. A correlation of the Sly Gap with the Devil's Gate formation of Nevada may also be suggested, at least tentatively.[59]

THE CONTADERO FORMATION

Contadero is a new formational name proposed for a series of carbonaceous shales and limestones above the Sly Gap formation and below the Mississippian strata in the central part of the San Andres Mountains. The type section is located in the south half of Sec. 8, T.-13-S., R.-4-E., near an abandoned mining claim, and in a small tributary draining part of the north slope of Rhodes Canyon. It can be approached on New Mexico Highway No. 52, which passes through Rhodes Pass between the towns of Tularosa and Hot Springs. The type section can be seen approximately 2,000 feet north of the road, and orientation is facilitated by three near-by prospect pits on the slope of the canyon. The name "Contadero" was taken from a Minor Civil Division Map, of the Bureau of Census, issued in 1934. The Contadero political division includes Rhodes Pass and the surrounding area.

The basal gray limestone beds of the Contadero rest without apparent disconformity on the Sly Gap formation, but the gray-black shales and thin, limestone top beds of the Contadero are separated from the overlying Caballero formation by an erosional disconformity (Fig. 12).

Because of the restricted outcrop area of the Contadero, correlations cannot be made to the east or west in the Sacra-

57 "Echinoderms of the Iowa Devonian," *Iowa Geol. Surv.*, Vol. X (1919–20) (annual reports), pp. 391–506.

58 A. K. Miller, personal communication.

59 C. W. Merriam, "Devonian Stratigraphy and Paleontology of the Roberts Mountains Region, Nevada," *Geological Society of America Special Paper No. 25* (1940).

THE SLY GAP FORMATION FAUNAL LIST

BRACHIOPODA
- *Athyris* cf. *vitata*
- *Atrypa rockfordensis* Fenton and Fenton
- *A. devoniana* Webster
- *Atrypa* n. sp.
- *Aulacella* sp.
- *Camarotoechia* cf. *duplicata*
- *Cranaena* cf. *calvani*
- *Cyrtina hamiltonensis* Hall
- *Cyrtospirifer whitney* (Hall)
- *Dalmanella?* sp.
- *Douvillina* cf. *arcuata*
- *Emanuella?* sp.
- *Gypidula cornuta* Fenton and Fenton
- *Hypothyridina* cf. *emmonsi*
- *Hypothyridina* n. sp.
- *Nervostrophia* sp.
- *Productella walcotti* Fenton and Fenton
- *Productella* sp. indet.
- *Pugnoides* cf. *calvani*
- *Schizophoria* sp.
- *Schuchertella* cf. *prava*
- *Spirifer* cf. *orestes*
- *Spirifer* cf. *disjunctus*
- *Stropheodonta inflexa* Swallow
- *Strophonella* sp.
- *Strophonelloides reversus* (Hall)
- *Tenticospirifer?* sp.
- *Tylothyris* cf. *mesacostalis*

CRINOIDEA
- *Arthroacantha* n. sp.
- Crinoid roots and columnals

ANTHOZOA
- *Alveolites rockfordensis* (Hall and Whitfield)
- *Aulopora* cf. *saxivadum*
- *Aulopora* sp.
- *Chonophyllum?* sp.
- *Coenites* cf. *floydensis*
- *Hexagonaria* sp.
- *Macgeea solitaria* (Hall and Whitfield)
- *Phillipsastrea* cf. *woodmani*
- *Tabulophyllum?* sp.

BRYOZOA
- *Fenestrellina* sp. indet.
- *Lioclema slygapensis* Fritz

CEPHALOPODA
- *Manticoceras regulare* Fenton
- *Orthoceras* sp.

GASTROPODA
- *Bellerophon* sp.
- *Platyceras* sp.
- *Straparollus* sp.

PISCES
- *Arthrodire* cf. *Coccosteus*
- *Deltodus*
- *Orodus*
- *Petalodus*
- *Ptyctodus*

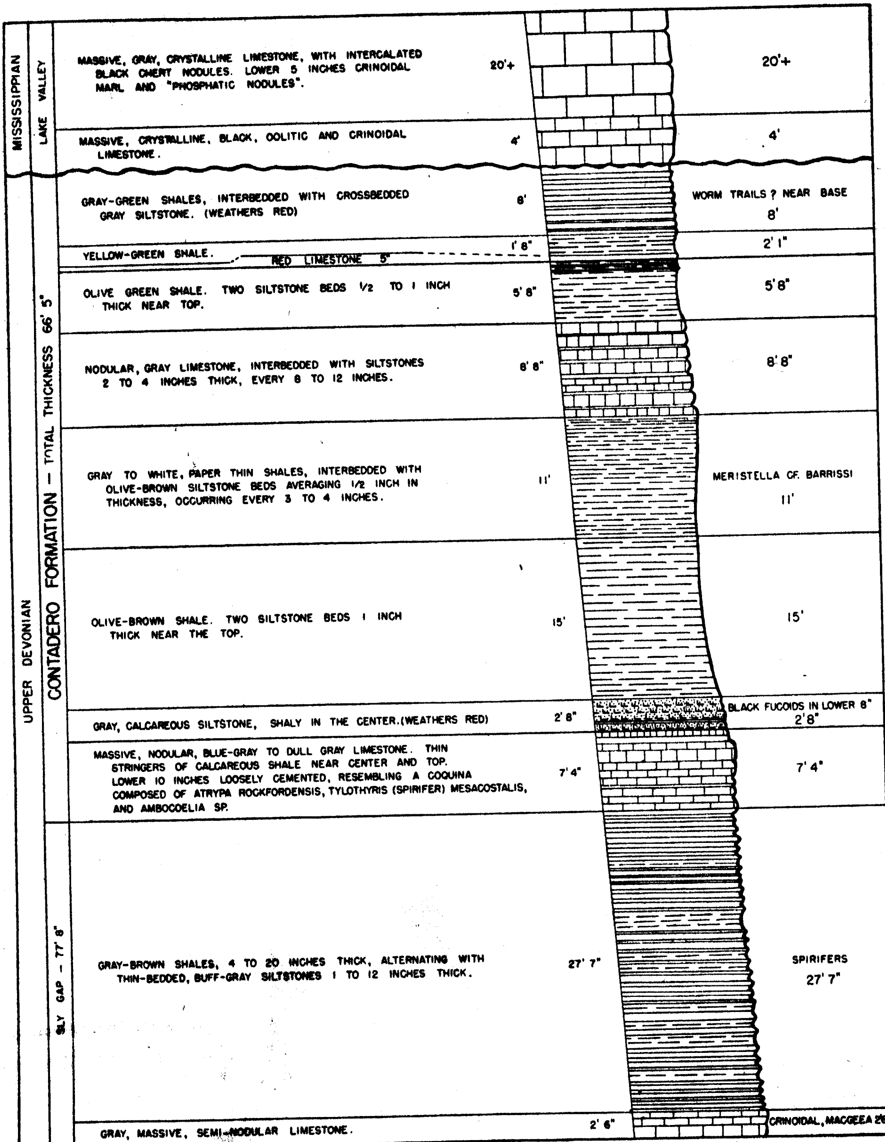


Fig. 12.—Type-section chart of the Contadero formation

mentos or Sierra Caballos. In cross section from north to south in the San Andres Mountains the Contadero formation roughly resembles the shape of a shallow bowl (Fig. 6).

Designation of the Contadero in this paper as a formation does not preclude the possibility of other stratigraphic terms eventually being applied to these beds. The Contadero formation may prove to be a tongue, member, or lentil of the Sly Gap formation; or it may be the age equivalent of the Ready Pay member of the Percha shale. Nevertheless, the name "Contadero" will retain its usefulness as designating a mappable and distinct lithologic unit, regardless of any subsequent change in its stratigraphic classification.

The fauna of the Contadero is, in general, the same as that of the Sly Gap, and undoubtedly many of the species are conspecific. An undescribed species of *Ambocoelia* which has not been found in the Sly Gap, however, occurs abundantly in the basal beds of the Contadero. In many basal sections the *Ambocoelia* are so abundant that the rock is essentially a coquina.

In the type of preservation of the fossils, however, the contrast seen in the two faunas is as great as the differences in the lithology of the Sly Gap and the Contadero. The Contadero fauna is preserved in gray argillaceous limestone with layers of white calcite representing the actual shell structure, whereas the Sly Gap fossils have a center of buff, dolomitic limestone and, although similarly layered with calcite, the dolomite matrix gives a yellow-buff color to the specimens.

THE PERCHA SHALE

The Percha shale is best exposed in the region between Lake Valley, Hillsboro, and Kingston and in the vicinity of Santa Rita and Silver City. There are also many outcrops of the Percha between these two regions. It was the only recognized Devonian formation in New Mexico from 1907 through 1940, inclusive (Fig. 2), although the presence of other Devonian formations was suspected.

The specific type section of the Percha shale has not been certainly identified. The type locality was designated by Gordon[60] as on Percha Creek in the vicinity of Hillsboro, New Mexico. In this area very few sections of the Percha shale expose the basal and upper contacts, but an excellent section (Fig. 13), which does show both contacts, is found 2½ miles southeast of Hillsboro in the SW.¼, SW.¼, SE.¼, of Sec. 14, T. 16 S., R. 7 W.[61] This exposure is approximately ½ mile south of the narrow canyon called "The Box," through which the Percha Creek flows eastward into the Rio Grande. This is the section that subsequent workers have probably referred to as the type, and the writer proposes that it be adopted as the neotype section. The area, however, is heavily mineralized, and the fossils occurring here are not so well preserved or so abundant as in other areas.

In 1942 the Percha shale was divided into the "lower Percha shale" and the "upper Percha shale"[62] (Fig. 2). But, because of possible future complications from this nomenclature, the writer proposes the name "Ready Pay member" for the "lower Percha shale" and "Box member" for the "upper Percha shale." The Ready Pay member was named after a gulch of that name which drains to the south into Percha Creek, 400

[60] Pp. 60 and 62 of ftn. 2 (1907).

[61] U.S. Geol. Surv., New Mexico Hillsboro Quadrangle (1940 ed.).

[62] Stevenson, pp. 23–24 of ftn. 18 (1942).

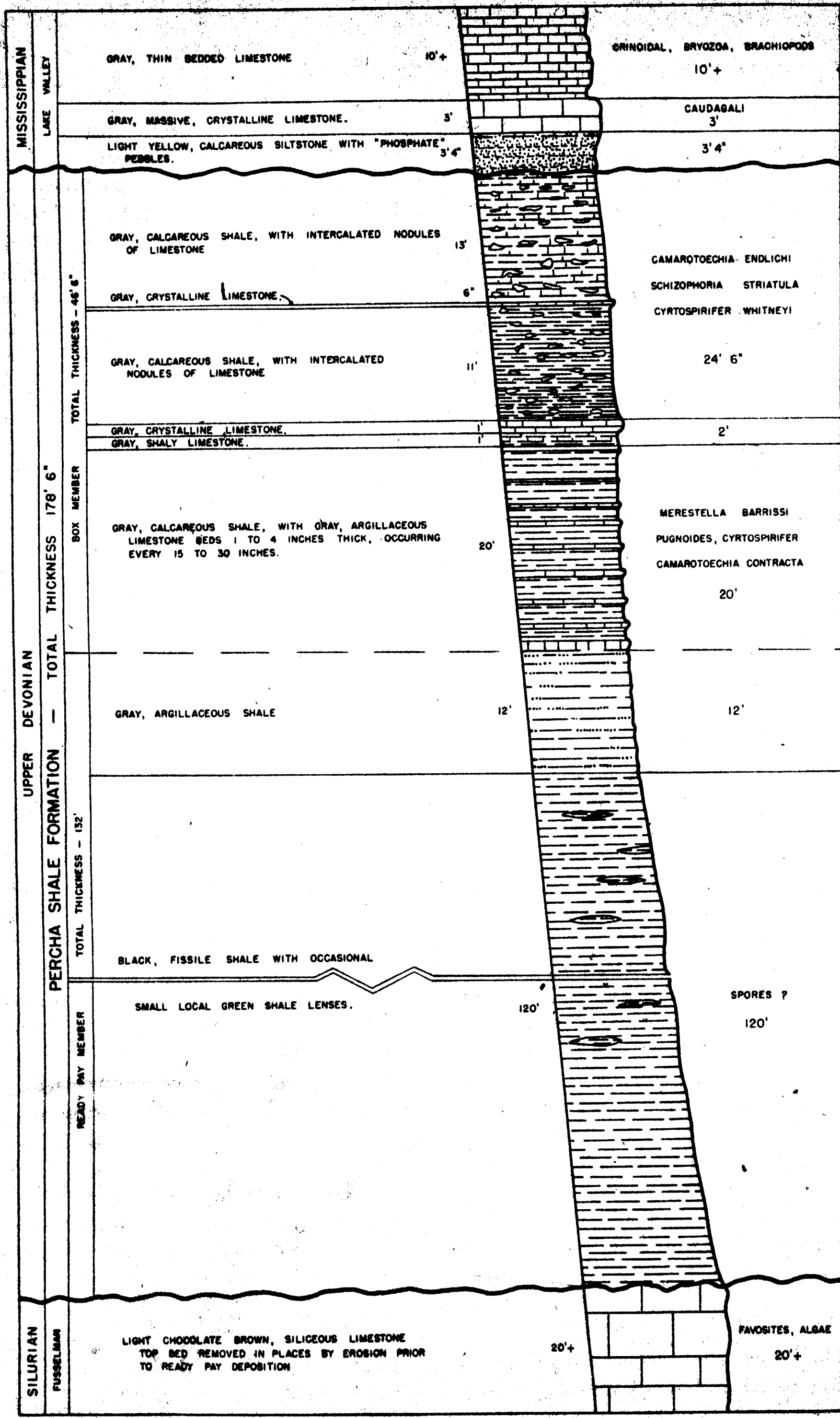


FIG. 13.—Type-section chart of the Percha shale formation. (The zigzag double line through the Ready Pay member represents shale not included to scale.)

yards east of "The Box," for which the upper member was named.

The Ready Pay member is composed of black, fissile, nonfossiliferous shale; and in most sections it comprises two-thirds of the total thickness of the Percha shale. The Ready Pay member grades without any marked break into the Box member. The Box member is composed of gray to green calcareous shales with intercalated lenses and nodules of limestone.

The Box member becomes progressively more calcareous to the west of Hillsboro; and on Bear Mountain, northwest of Silver City, interbedded in the shale are many massive beds of limestone, one attaining a maximum of 16 feet. This contrasts strongly with the scattered limestone nodules found in the Box member in the region of Hillsboro and Lake Valley. The Box member, which has not been recognized outside of the Mimbres region, carries the entire Percha fauna.

The Ready Pay member is believed to occur throughout southern New Mexico and probably in the Franklin Mountains of Texas. Black, fissile, nonfossiliferous shale has been found above the Sly Gap and below the Mississippian formations in the Sacramento and San Andres mountains. As previously pointed out, the Contadero formation, whose upper portion consists mainly of shales, has, in a few localities in the San Andres, supplanted the Ready Pay (Fig. 6). In Mud Springs and Caballos mountains the upper part of the Sly Gap becomes increasingly shaly, with black shale predominant at the top. In Cooks Peak northeast of Deming, New Mexico, 182 feet of black fissile shale is found below the Caballero formation. The Box member is recognized a short distance to the north but has not been detected in Cooks Peak. In the Franklin Mountains of Texas the Canutillo formation has approximately 40 feet of black, fissile, nonfossiliferous shale at the top and is overlain by the Helms formation (Mississippian). The shale may or may not be stratigraphically equivalent to the Ready Pay member.

There is no definite proof that the widely occurring black shales of southern New Mexico are correlatives of the Ready Pay, but the stratigraphic position of these shales and the general Paleozoic history of the region indicate their probable contemporaneity. This view is strengthened by the discovery of a relatively thin bed of brown, nonfossiliferous siltstone at the base of the Ready Pay in three widely scattered localities in the Mimbres region. Four feet of siltstone is found at the base of the Percha shale on North Percha Creek, near an abandoned Forest Ranger station. A sample of the silt was thin-sectioned (Fig. 7, *d*); and, when compared with the thin section from the "Spirifer zone" of the Sly Gap, very few differences were noted except for grain size. A similar occurrence was studied at the base of the Percha shale in Bear Mountain, where approximately 6 feet of brown siltstone is found. Bowsher[63] has also reported approximately 4 feet of brown siltstone below the Ready Pay member in Decker Draw, due east of P. A. Mountain, in the Hillsboro quadrangle.

Until a fauna is found in the Ready Pay or the brown siltstone underlying the Percha shale, or a new locality reveals the actual succession of formations, equivalency of the black shales will be difficult to establish. The brown siltstones can conceivably be either Canutillo, Onate, or Sly Gap in age, although the latter age is most probable. Possibly

[63] Personal communication.

the final solution of this problem will be through the use of well-log data. In the Rattlesnake Field in northwest New Mexico a deep test well penetrated 140 feet of green shale, dense tan limestone, and dense brown dolomite, believed to be Devonian in age.[64] In the event of deep wells being drilled in southern New Mexico, well-log data will unquestionably permit more positive surface identifications, as well as correlations with subsurface strata in areas outside of southern New Mexico.

The Percha shale fauna was first studied by E. M. Kindle;[65] and M. A. Fritz[66] has recently described eight new species of bryozoans from the Box member, which is undoubtedly the age equivalent of the Ouray limestone of Colorado. Collections from the Box member of the Percha shale made by the writer indicate that approximately 80 per cent of the fauna has been described. Revision of the described forms and description of the remaining unidentified fossils will be undertaken sometime in the future.

CONCLUSIONS

There are many lines of evidence which must be more thoroughly explored before a completely satisfactory picture of the Devonian of New Mexico can be presented.

The Onate formation is difficult to interpret. The lack of a diagnostic faunal assemblage at present precludes definite correlations with formations to the east or west. The exact age of the Onate may finally fall somewhere within the time range from the Onondagan to the late Hamilton.

A rather complete picture of the Sly Gap formation is presented in southern New Mexico, but its relationships with the Percha shales to the west are unknown. On the basis of faunal evidence, the Sly Gap is definitely distinct from the Percha fauna and that of the Ouray limestone. Only long-ranging genera and species have been found both in the Percha and Sly Gap assemblages, and only a relatively low percentage of these. Through the study of large Devonian collections in Walker Museum, it was found that more Sly Gap forms were identical with species from the Hackberry of Iowa than with those from any other formation. There are several differences between the two faunas, and the Sly Gap is probably older than the Hackberry. The Sly Gap fauna is probably younger than the fauna of the Independence shale of Iowa, inasmuch as several forms that appear in the Sly Gap are only found stratigraphically higher than the Independence. If these conclusions concerning the Sly Gap are correct, it is older than the Martin limestone of Arizona, which is definitely a time equivalent of the Hackberry. According to A. A. Stoyanow,[67] the stratigraphic position of the Martin limestone is lower than the Lower Ouray formation on Pinal Creek, north of Globe, Arizona. The "Lower Ouray" designation has been employed as a working term by Stoyanow to indicate sediments containing the fauna found in the Ouray limestone of Colorado, which is undoubtedly an equivalent of the Box member of the Percha shale. Although there is no known sec-

64 "Rattlesnake Field: The Oil and Gas Resources of New Mexico," *N.M. School of Mines, State Bur. Min. and Min. Res. Bull. No. 18* (2d ed., 1942), p. 119.

65 Ftn. 3 (1909).

66 Pp. 31–41 of ftn. 19 (1944).

67 Pp. 489–93 of ftn. 7 (1936).

tion exposing the Sly Gap below the Percha shale, any correlation of the New Mexico formations with those of Arizona brings out the fact that the Sly Gap formation is older than the Percha.

Devonian specimens figured by C. W. Merriam,[68] from the "Cyrtospirifer zone" of the Devils Gate formation, Nevada, resemble many forms in the Sly Gap fauna; and a correlation of time-stratigraphic equivalency of the two formations is likely.

The geologic history of southern New Mexico during late Devonian and Mississippian times seems to indicate low, unstable land masses and fluctuating seas. There is the possibility that the Sly Gap formation may be represented by three facies: (1) the lower fossiliferous, typical Sly Gap sediments; (2) the Contadero formation as a lentil; and (3) the Ready-Pay-member type of strata, possibly representing deposition in a low marshy area on the fringes of the main Sly Gap sea (Fig. 10). Sedimentation under such conditions would in part explain the failure to find the Sly Gap fauna in the area in which the Box member of the Percha shale occurs.

[68] Ftn. 67.

Acknowledgments.—Grateful acknowledgments are due to Lieutenant A. L. Bowsher, of the U.S. Army, Dr. L. R. Laudon, of the University of Kansas, Dr. J Harlen Bretz, of the University of Chicago, and Dr. L. A. Nelson, of the El Paso School of Mines, for valuable field assistance. My sincere thanks are due to Mr. Lawson P. Entwistle, of the Ground Hog Mine, Vanadium, New Mexico, Dr. A. K. Miller, of the State University of Iowa, and Dr. G. A. Cooper, of the National Museum, for aid in the stratigraphic and paleontologic sections of this paper, and to Major H. G. Schenck, of the U.S. Army, for helpful suggestions concerning stratigraphic terminology.

Thanks are also extended to many residents of southern New Mexico, whose hospitality and interest in the writer's work have been of inestimable aid.

Appreciation for many helpful comments and suggestions is expressed to Dr. Carey Croneis, under whose direction this paper was written, and to other members of the staff at the University of Chicago. Large collections of Devonian fossils on deposit at Walker Museum, University of Chicago, and the collections of Sly Gap fossils on loan from Dr. L. A. Nelson and the Agricultural and Mechanical College of Texas, Lubbock, Texas, have enabled the writer to form many conclusions incorporated in this paper.

ORIENTATION ANALYSIS OF FINE-GRAINED CLASTIC SEDIMENTS: A REPORT OF PROGRESS

E. C. DAPPLES AND J. F. ROMINGER
Northwestern University

ABSTRACT

The purpose of this investigation is to extend methods of orientation analysis to fine-grained clastic sediments in order to determine the character of their last depositional environment, including the nature and direction of movement of the depositing agent. A two-dimensional method, based upon statistical analysis of measurements of the projections of grains on the bedding plane, is developed. Two measurements are made on each grain projection: (1) elongation azimuth and (2) end position (defined as the position of the larger end of the grain with respect to the smaller end). Grains from laboratory fluvial and eolian environments exhibit a pronounced preferred elongation parallel to the direction of flow of the depositing agent and a marked tendency to lie with their larger ends upcurrent. Results of this study indicate that the source direction of fine-grained fluvial and eolian sediments can be determined uniquely by use of the two measurements and suggest that the agent of deposition of a sediment may be determinable from its orientation pattern.

INTRODUCTION

Investigators concerned with the primary structure of sedimentary rocks must attach certain significance to the mineral composition, shape, surface texture, size, and arrangement of the component grains. The degree that these physical properties characterize the environmental history of a sediment, however, is a matter of considerable uncertainty, arising largely from an inability to evaluate the observed data properly. Interpretation is made difficult by the complexity of these properties, which results from differences in the number, the succession, and the nature of the environments to which each mineral grain may be subjected. Collectively, the visible effects of the environments too often tend to confuse rather than to facilitate reconstruction of the depositional history of a sediment, and in such instances it becomes necessary to interpret the last phase of the history on the basis of certain selected physical properties. Of these, the arrangement of mineral grains in unmetamorphosed sediments appears to be least influenced by environments previous to the last and hence should provide one of the most reliable bases for interpretation of the ultimate depositional environment of a sediment.

The existence of preferred dimensional orientation of pebbles in coarse-grained unconsolidated deposits has been demonstrated by Hakon Wadell,[1] Konrad Richter,[2] and W. C. Krumbein.[3] It is reasonable to expect, therefore, that orientation patterns exist in fine-grained clastic materials. If such patterns remain fixed in a sediment after deposition and show tendencies to be diagnostic of particular environments of deposition, they should provide a means of deter-

[1] "Volume, Shape and Shape-Position of Rock Fragments in Open-Work Gravel," *Geografiska Ann.*, Band XVIII (1936), pp. 74–92.

[2] "Die Bewegungsrichtung des Inlandeises, rekonstruiert aus den Kritzen und Längsachsen der Geschiebe," *Zeitschr. f. Geschiebeforschung*, Band VIII (1932), pp. 62–66; "Gefüge und Zusammensetzung des Norddeutschen Jungmoränengebietes," *Greifswald Univ., Geol.-Pal. Inst. Abhand.*, Heft XI (1933), pp. 1–63.

[3] "Preferred Orientation of Pebbles in Sedimentary Deposits," *Jour. Geol.*, Vol. XLVII (1939), pp. 673–706; "Flood Deposits of Arroyo Seco, Los Angeles County, Calif.," *Bull. Geol. Soc. Amer.*, Vol. LIII (1942), pp. 1386–91.

mining the conditions under which a sediment was last deposited.

Wadell[4] and Krumbein[5] have developed methods by which the position in space of an individual pebble may be defined. By these methods data are accumulated on the positions in space of many coarse particles and analyzed statistically to determine the existence of preferred dimensional orientations. However, application of the procedure of Wadell and Krumbein is limited by difficulties encountered when the particles are too small to handle and measure individually.

It is the purpose of this investigation to develop a method of orientation analysis for fine-grained particulate materials, to investigate the existence of diagnostic orientation patterns in fine-grained sediments, and to establish principles concerning such patterns. If the quantitatively predominant fine-grained clastic sediments can be analyzed in this manner, the results obtained should prove valuable in the solution of such problems as (1) determination of the direction of the source of a sediment, (2) determination of the agent of deposition of a sediment, and (3) correlation among sedimentary formations.

At present the authors wish to report a method of determining the orientation of the coarser elements of fine-grained particulate materials. Preliminary results employing this method reveal the presence of what may prove to be diagnostic orientation patterns in fine-grained sediments; at least such patterns were discovered in the laboratory sediments examined. Because of the restricted conditions under which the sediments were deposited in the laboratory, the results are limited in scope, and conclusions of a general nature must await more intensive and extensive research.

TERMINOLOGY

As used herein, the term "preferred orientation" refers to a spatial arrangement of grains which can be shown by accepted statistical methods to exhibit a significant deviation from random disposition (i. e., from disposition with an equal number of grains in each possible position). Orientation is considered only from a geometrical standpoint—the dimensional orientation of E. B. Knopf and Earl Ingerson.[6] Since no crystallographic directions are employed, the term "fabric," used by these authors to signify "not only the pattern produced by the external shape of rock constituents, but also the pattern of the internal elements down to the very space lattice of the rock minerals,"[7] is not applicable. Consequently, the term "dimensional orientation pattern," or simply "orientation pattern," understood to refer to the summation of the dimensional orientations of the component grains of a material, is used throughout this discussion. Data concerning dimensional orientation patterns would have to be combined with all other structural and textural data to describe the fabric of any material; such an integration is beyond the scope of this paper.

New terms used in this study are shown by italics.

CLASSIFICATION OF DIMENSIONAL ORIENTATION PATTERNS

To facilitate discussion of experimental data, it is desirable to establish a terminology by which the various types of dimensional orientation patterns may be

[4] Ftn. 1 (1936).

[5] Ftn. 3 (1939).

[6] "Structural Petrology," *Geol. Soc. Amer., Mem. VI* (1938), p. 10.

[7] *Ibid.*, p. 12.

described. Classification by agent is immediately suggested and affords such descriptive terms as "eolian," "aqueous" (fluvial, littoral, marine, lacustrine), and "gravity" or "angle of repose" (subaerial, subaqueous) dimensional orientation patterns. In addition, division of each of these agent *classes* into the following *groups* is helpful.

1. *Translational orientation pattern.*—Dimensional orientation pattern of material while it is in the transporting medium and before deposition.[8] This group may be subdivided into (*a*) suspension, (*b*) saltation, and (*c*) traction *phases* or into (*a*) bed-load and (*b*) suspended-load phases.[9]

2. *Depositional orientation pattern.*—Dimensional orientation pattern resulting from deposition and continued action of the transporting agent after translatory motion of the particles has ceased. This pattern must be developed during deposition and in the period when the deposit is in contact with the transporting medium. It may vary with the length of time between deposition and burial.

3. *Secondary orientation pattern.*—Dimensional orientation pattern resulting from any factors (natural or artificial) other than the agent of deposition. Phases of this group would include patterns due to weathering, metamorphism, effects of organisms, etc.

The above method of classification has the advantage of being complete and of using terms that are concise and suggestive of the nature of the orientation pattern. Orientation patterns may now be described uniquely by the use of combinations of the terms suggested—for example, translational aqueous orientation pattern, translational fluvial orientation pattern (saltation phase), depositional marine orientation pattern, and secondary recrystallization orientation pattern. A more complicated terminology may be required to describe results of advanced research, but this system is adequate for present purposes.

METHOD OF INVESTIGATION

The complete analysis of the position of a particle in a stratum is a three-dimensional problem and should include the determination of both azimuth and angle of dip of the grains. In describing the orientation of coarse clastic sediments, Krumbein[10] presents methods which satisfy these requirements. With fine-grained sediments, however, manipulation of the particles becomes difficult, precluding the application of Krumbein's methods. No satisfactory means of handling the small grains individually and no other successful three-dimensional method of analysis was discovered in this investigation. Consequently, the writers have investigated the possibilities of using a less complete system of two-dimensional analysis which does not involve removal of the particles from their original position. This method, depending primarily on the outline shape of the projection of grains, apparently provides sufficient diagnostic data to merit use.

A specimen of the material whose orientation pattern is to be investigated is photographed through a standard petrographic microscope. Magnification is increased further by photographic enlargement until the dimensions of the grains

[8] The translational behavior of nonspherical particles has been investigated by Krumbein, "Settling-Velocity and Flume-Behavior of Non-spherical Particles" *Trans. Amer. Geophys. Union, 1942*, Part II (1942), pp. 621–33; see also Wadell, "Some New Sedimentation Formulas," *Jour. App. Phys.*, Vol. V (1934), pp. 281–91.

[9] H. A. Einstein, A. G. Anderson, and J. W. Johnson, "A Distinction between Bed-Load and Suspended-Load in Natural Streams," *Trans. Amer. Geophys. Union, 1940*, Part II (1940), pp. 628–33.

[10] Ftn. 3 (1939).

are raised to lengths ranging from $\frac{1}{2}$ to 1 inch. Best results are obtained by using low-power magnification with the microscope and photographing onto a $3\frac{1}{4}'' \times 4\frac{1}{4}''$ negative. In this manner a large number of grains in sharp focus is obtained, and magnification is achieved principally through enlargement of the negative. Correct orientation of each photograph is insured by suspending a needle point or finely drawn glass capillary over the northeast quadrant of the microscope field.[11]

The method of determining and statistically studying the grain orientation is based upon the projection of the grains visible on the stratification plane. The bedding plane was selected as a surface of prime importance for several reasons: (1) it is generally recognized easily in outcropping strata and in drill cores; (2) knowledge of the orientation of the grains in the stratification-plane direction is necessary to give the source direction of the sediment; (3) the surface approaches a plane; (4) slight weathering of the bedding surface commonly results in the grains' standing out in relief; hence, in the photograph the grain boundaries are distinct.

The authors are of the opinion that it may be possible to determine the true orientation of a grain from its projections on three mutually perpendicular planes. However, calculations based upon this assumption, considering the grains to be triaxial ellipsoids, have yielded results expressible only in mathematical terms of an order too high to permit practical usage; whereas the two-dimensional method, using the projection of grains on the bedding surface, has as its greatest advantage the simplicity of calculation and, as far as the investigation has been carried, does reveal the nature of dimensional orientation patterns.

A practical approach to three-dimensional analysis would be to use projections on two additional planes, one parallel and one perpendicular to the direction of movement of the grains at the time of deposition (as determined by analysis of the bedding-plane orientation pattern) and both perpendicular to the bedding plane. This method should disclose imbricate or edgewise structures and, being more complete, probably would permit reliable interpretation of depositional environments in some cases where bedding-plane analysis alone would be inadequate. The study of this method should await development of the bedding-plane method of determining current direction, upon which it depends, and which in itself yields one of the major results desired. It is the authors' intention to investigate this more nearly complete three-dimensional method at a later date.

The nature of an orientation pattern is revealed only by a statistical analysis of the attitude of many grains. Study of the positions of individual grains is of little significance in a system so complicated by irregular grain shapes and wide range of orientation. For the collective study necessary in such an analysis, statistical methods provide means of synthesizing the data, and criteria for judging whether the observed orientation distributions are due to chance alone or follow some other law of distribution (i.e., whether they exhibit random or preferred orientation).[12]

[11] With further projection-orientation studies, greater rapidity of analysis with satisfactory results may be obtainable by use of a microprojector. See W. D. Pye, "Rapid Methods of Making Sedimentational Analyses of Arenaceous Sediments," *Jour. Sed. Pet.*, Vol. XIII (1943), pp. 86–88.

[12] The statistical notation used herein is that of J. F. Kenny, *Mathematics of Statistics*, Parts I and II (New York: D. Van Nostrand Co., 1939).

PROBABILITY THEORY

In this investigation it has been necessary to determine whether the distributions obtained are due to natural factors or represent merely chance deviations from random distributions. For this purpose the theory of the sampling of attributes has been employed.[13] By this method it is possible, for example, to determine the probability of occurrence of sixty heads and forty tails in throwing a hundred coins. If the calculated probability in such cases is less than 0.01 (1 chance in 100), a conservative limit, the deviation from chance is said to be significant, and it may be concluded that the coins are biased. On the other hand, if the probability is greater than 0.05 (1 chance in 20), the deviation from the expected value of fifty heads and fifty tails is not regarded as significant and may be accepted as due to chance fluctuations. It should be noted that the 0.01 and 0.05 limits are arbitrary values that have been adopted by statisticians as a result of experience; in the range 0.01–0.05 no conclusions can be drawn with assurance. Tests according to the sampling-of-attributes theory may be made to evaluate the significance of the deviation of any variable from its expected value.

Each distribution in this study has been tested to determine whether or not it exhibits a significant deviation from the value expected by chance alone. If the calculation indicates that the deviation is significant, it is concluded that the factor other than chance is a physical process of sedimentation.

According to the sampling-of-attributes theory, if an event has the probability of occurence p, the probability that the number of occurrences x in s trials will differ from the expected number $E = sp$ by as much as d is given approximately by $P_\delta = 1 - Q_\delta$, where

$$Q_\delta = 2\int_0^\delta \phi(t)\,dt\,,$$

$$\delta = \frac{d - \frac{1}{2}}{\sigma}\,,$$

$$\phi(t) = (2\pi)^{-\frac{1}{2}} e^{-\frac{t^2}{2}}\,,$$ the equation of the normal probability curve,

$$\sigma = (spq)^{\frac{1}{2}}\,,$$ the standard deviation,

$$q = 1 - p\,,$$ the probability of non-occurrence.

It will be noted that the integral in the equation for Q is that evaluated in mathematical tables as "area under the normal probability curve." The probability calculations used in this study are all made according to these equations.

TYPES OF MEASUREMENTS

Two main types of measurements are employed in this analysis: (1) elongation direction (p. 251) and (2) end position (p. 251) with respect to that direction. Both elongation direction and end position are included in the concept of preferred dimensional orientation, and, as demonstrated below, a combination of these two dimensions is necessary to determine uniquely the direction of motion of the depositing agent of a sediment. Both of these dimensions might well be included in Wadell's[14] term "shape-position."

ELONGATION DIRECTION

Three methods of measuring the elongation direction of grain projections, each with accompanying advantages and disadvantages, are recognized in the present study.

[13] *Ibid.*, Part II, pp. 24–26.

[14] Ftn. 1 (1936).

1. *Long-dimension elongation.*—Direction of elongation is considered as that of the longest line which can be drawn on the grain projection (Fig. 1). This dimension is the projection equivalent of the three-dimensional measure of elongation used by Krumbein[15] for coarse clastic sediments. It has the advantage of ease of determination but loses value by its susceptibility to irregularities of the grain projections.

2. *Least-projection elongation.*—Elongation direction is considered as that of the two parallel lines with the minimum amount of separation which can be drawn tangent to the grain projection (Fig. 1). In practice these lines are constructed readily with the aid of a parallel rule. This dimension has theoretical importance in that the least projection in most cases probably offers least resistance to a circumfluent medium.

3. *Center-of-area elongation.*—Elongation direction is considered as that of the longest straight line which can be drawn through the center of area of the grain projection. (The term *center of area* is defined herein as the point of intersection of all lines bisecting the area of a grain projection. It is, therefore, the two-dimensional equivalent of center of mass.) This method has the disadvantage of requiring the use of a planimeter on each grain projection. However, it probably enjoys the soundest theoretical basis of the three methods, since the center of area of a grain projection is in most cases nearly above the center of mass of the grain and, hence, above the point about which the grain pivots when suspended in a fluid.

END POSITION

Knowledge of preferred elongation direction, and the fact, as demonstrated below, that this direction is parallel to the current direction in fluvial and eolian environments, limits the current to two possible directions. For example, if the elongation trend of grains in a fluvial environment is found to be statistically north-south, the current could have flowed either from the north or from the south. A method of making the correct choice between the two possibilities is provided by determining the position of

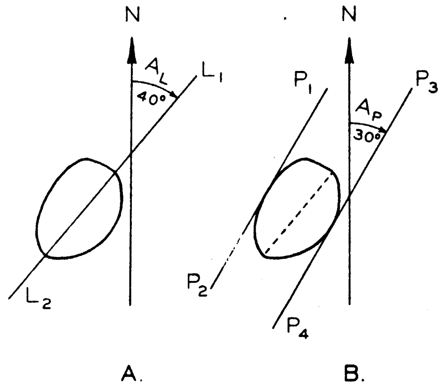


FIG. 1.—Comparison of long dimension and least projection measures of elongation azimuth. *A*, Grain with long dimension axis (L_1L_2) making an azimuth angle (A_L) of 40° from north (N). *B*, Same grain with least projection tangents (P_1P_2, P_3P_4) showing a least projection azimuth (A_P) of 30°.

the larger end of the grain with respect to the smaller end, herein defined as *end position*, for each grain, in terms of the two opposite directions indicated by the elongation analysis (north and south in the example just cited). Use of this measurement rests on the assumption that the larger end of a grain tends to be upcurrent from the smaller end; a relation known to exist in falling drops of water. In this position, corresponding to the so-called "tear-drop" streamlined shape, resistance to flow is at a minimum.

End position, as defined, is a three-dimensional measurement—a compari-

[15] P. 677 of ftn. 3 (1939).

son of the relative volumes of the two parts of a grain delimited by the plane bisecting its elongation axis, and perpendicular to that axis. As such, it may not be the exact property desired; in fact, even the least-resistance principle hypothesized may not be valid. However, the results obtained by use of one of the following measurements, each of which represents a projection approach to the determination of the three-dimensional quantity, make use of the concept of end position valuable.

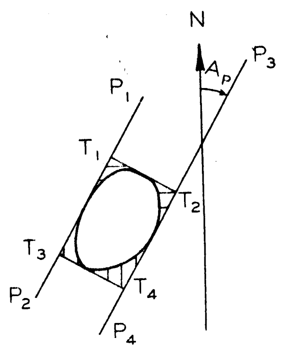


FIG. 2.—Least projection rectangle ($T_1T_2T_3T_4$) constructed about Figure 1 grain projection by erecting tangents (T_1T_2, T_3T_4) perpendicular to the least projection tangents (P_1P_2, P_3P_4). The smaller area (horizontally ruled) outside of the grain and within the least projection rectangle is at the north (N) end of the rectangle. Hence the greater area is in the north portion of the grain—i.e., the area direction is north.

Three methods of describing end position, one corresponding to each of the elongation measurements, afford possible quantitative means of distinguishing the larger end of a grain from its two-dimensional projection.

1. Intersection of axes position: In conjunction with the long-dimension axis, and corresponding essentially to the intermediate axis of Krumbein,[16] is the longest line that can be drawn on a grain projection perpendicular to its long-dimension direction. If it is not affected too much by irregularities in the grain outline, the position of the intersection of this line on the long-dimension axis, with respect to the center of that axis, should in most instances be indicative of the end position of the grain.

2. Area direction: The position of the greater area of a grain projection with respect to the center of its elongation axis, herein defined as *area direction*, is evidently the projection equivalent of end position. The precise method of determining this dimension is by using a planimeter to measure the area on either side of the perpendicular bisector of the elongation axis. As such a process is somewhat laborious, the authors have devised a method, requiring only visual inspection, which provides a reliable measure of end position. In this method the *least-projection rectangle* (rectangle formed by the least-projection tangents to the grain projection and the perpendicular pair of tangents) is constructed around each grain. The end of the grain having the greater area may then be determined by visual estimation of the half of the least-projection rectangle having the least area outside of the grain projection (Fig. 2). With a little practice the area direction of most grains may be determined with confidence by this method. Grains to which the visual method cannot be applied with certainty may be omitted without harm, the omission having the effect of reducing the importance, in the total sample, of minor irregularities on nearly symmetrical grains.

3. Center-of-area position: The position of the center of area of a grain with respect to the midpoint of the center-of-area elongation line is a measure of the grain's end position. As previously stated, the procedure for determining the center of area is lengthy, and, hence,

[16] *Ibid.*

determination of end position by this method is not accomplished with the rapidity of methods (1) and (2).

METHOD EMPLOYED

The measures of elongation, coupled with the corresponding means of determining end position, give three complete methods of analysis: (1) long-dimension elongation—intersection of axis position; (2) least-projection elongation—area direction; and (3) center-of-area elongation—center-of-area position. It is significant that both the theoretical soundness and the practical difficulty increase from method (1) to method (3). In view of this fact the authors have selected method (2) as the one which embodies the optimum combination of theoretical accuracy and operational facility. In the case of method (1) an orientation of the dimensional elongation of the grain projection is revealed which closely approaches the azimuth of current direction. On the other hand, this method fails to demonstrate that a relationship exists between the end position and the direction of current flow of the transporting medium (p. 255).

This weakness is overcome by method (2), which not only reveals a more decided preferred elongation (p. 255) but also gives a diagnostic measure of end position (p. 255). As method (2) yields satisfactory results, no study of the more difficult method (3) has yet been undertaken. It is possible, however, that depositional orientation patterns for sediments deposited in environments not investigated in this study may be revealed by method (3) and not by method (2). Until such proves to be the case, the advantages of ease and speed of determination of the least-projection–area-direction method should not be sacrificed.

ANALYSIS OF LABORATORY FLUVIAL SAND

The concepts discussed above arose from, or are based upon, an investigation conducted (1) to determine whether orientation patterns could be discovered in sediments deposited in the laboratory and, if so, (2) to determine the relationship between the pattern and the direction of the current which carried the component particles. The investigation included a study of the orientation of sand grains deposited under conditions which would closely simulate those in fluvial, eolian, and subaerial gravity (angle of repose) environments.

The apparatus used to produce the fluvial environment was a stream table 4 feet long, $3\frac{1}{2}$ inches wide, and $1\frac{1}{4}$ inches deep, constructed in a series of removable sections. Sand deposited on any part of the table could in this way be removed and studied without disturbing the grains. From a well at one end of the table, water was made to flow as a sheet slightly more than $\frac{1}{2}$ inch thick, covering the entire width of the trough. The inclination of the table was equivalent to a gradient of 90 feet per mile. Unsieved sand from the Lake Michigan beach along the Northwestern University campus[17] was poured into the upper end of the trough at a rate sufficient to maintain a small ridge of sand over which the water passed and gathered its load. Despite the uniform flow from the well, several rather strong cross-currents developed and could not be eliminated. Arbitrarily designating the source direction as due north, the cross-currents flowed

[17] Median diameter 1/10 mm. For a mechanical analysis of this sand see J. P. Todd, "Survey of the Near-Shore Deposits of Lake Michigan Adjacent to Northwestern University" (unpublished Master's thesis, Northwestern University, 1937), p. 32.

from northeast to southwest and from northwest to southeast. After ½ inch of sand had been deposited in the north (upstream) end of the trough, the flow of water was stopped and the sand allowed to dry. The section of the table from 4 to 16 inches below the sand source was then removed and placed under a petrographic microscope, the axis of the trough (north-south direction) being kept parallel to one of the cross-hairs. Twelve photographs, beginning at a point 8 inches below the sand source and continuing down the axis at ½-inch intervals, were taken through the microscope. It should be noted that the photographs constitute grain projections in a direction equivalent to the stratification plane. The first four pictures were of the upstream portion that apparently was unaffected by cross-currents, whereas the remaining ones were in the cross-current area. All the grains appearing distinctly on the enlarged pictures were then outlined in pencil, the pictures and grains numbered, and north lines drawn on each picture. On each of 507 grain projections thus delimited, the following measurements were made:

1. Azimuth angle of long-dimension axis: The long-dimension axis was first drawn on each grain projection, its position being determined accurately by the use of dividers. The azimuth of this axis, measured clockwise 0°–180° from north was then determined to the nearest degree for each grain. This is a measure of long-dimension elongation.

2. Azimuth angle of least-projection axis: Least-projection tangents were constructed about each grain with the aid of a parallel rule. Azimuth angles were measured in the same manner as those of the long-dimension axis and recorded as a measure of least-projection elongation.

3. Position of intersection of axes: The longest possible line perpendicular to the long-dimension axis was constructed on each grain projection. The position of the intersection of this line with the long-dimension axis was then determined as north or south of the midpoint of the axis. This position was recorded as either north or south for all grains having elongation azimuths within 89.5° of the current direction.

4. Area direction: Least-projection rectangles were completed about each grain and the area directions determined visually by estimating the half of the rectangle having the least area outside the grain. The area directions of 447 of the 507 grains were determined, the remaining grains being too nearly symmetrical for reliable estimation by this method.

SUMMARY OF FLUVIAL ELONGATION AZIMUTH DATA

Compilation of azimuth data reveals a pronounced preferred elongation parallel to the direction of movement of the depositing agent. Azimuth distribution histograms of both long-dimension and least-projection directions (Figs. 3 and 4) show a distinct mode of azimuths within 10° of the current direction. In the least projection distribution the 0°–9° and 170°–179° classes (those within 10° of the current direction) each contain 51 grains, an excess of 23 over the expected number (507/18) if the grains were oriented at random. The probability of obtaining such a concentration by chance alone is 0.00000 (less than 1 in 200,000), a value far less than the 0.01 accepted by statisticians to measure the greatest allowable deviation from any acceptable theory of distribution.[18] These

[18] The 0.00000 probability is obtained by the method described on pp. 250–54, the actual cal-

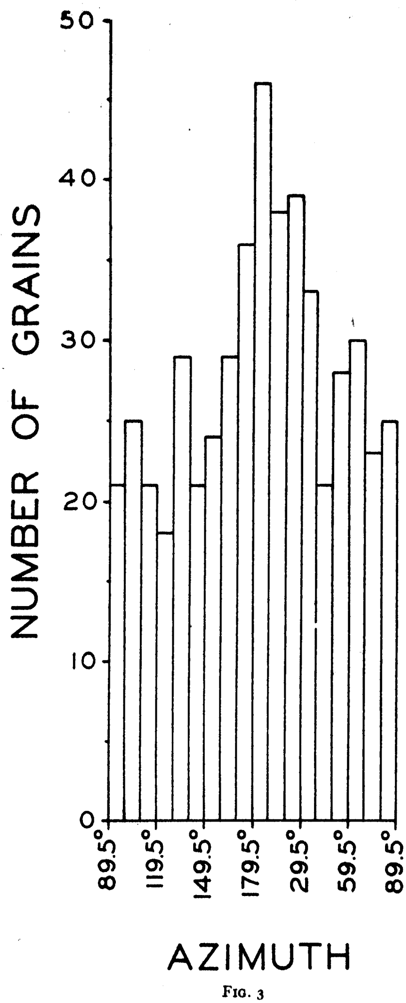


FIG. 3

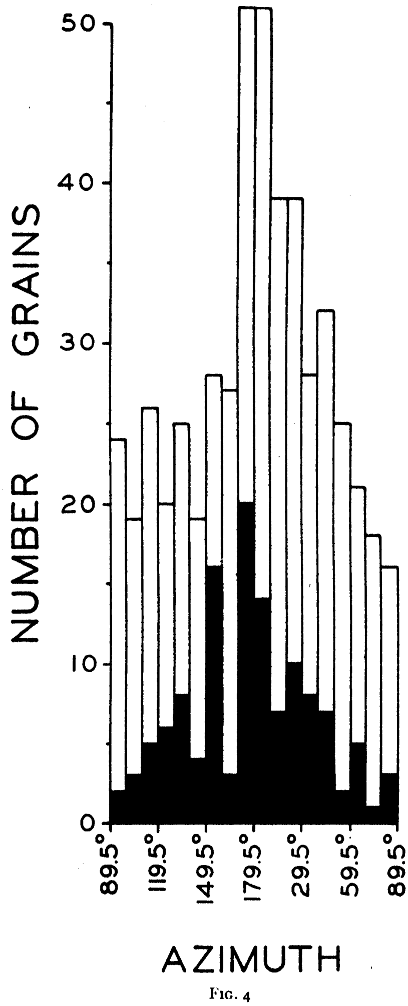


FIG. 4

FIG. 3.—(*Left*) Histogram of long dimension azimuth distribution of laboratory fluvial sand. Current direction = 0.0° (=180.0°). (Class boundaries, 89.5°, 99.5°, etc., are shown for classes 90°–99°, 100°–109°, etc.)

FIG. 4.—(*Right*) Histogram of least projection azimuth distribution of laboratory fluvial sand, showing distribution of grains in area with cross-currents (*white*) and in area without visible cross-currents (*black*). Current direction at source =0.0° (= 180.0°).

data constitute unquestionable evidence that grain elongation is not random but controlled by the depositing agent.

A statistical study of the two distributions of azimuths measured 0°–180° from the known current direction yields the characterizing constants shown in Table 1. It is evident that the current direction indicated by the mean of the azimuths in either type of measurement differs from the actual current direction (0.0°) by an insignificant amount. The skewness constants indicate a slightly greater number of azimuths in the 90°–180° than the 0°–90° range, a feature attributed to the effect of the irregularity of the strong cross-currents. The fact that the kurtosis values are less than 3.0 indicates that the distribution curves are slightly "flatter" than the normal probability curve.

culation being given here as an example of the method used throughout this paper.

Since, by chance alone there are 18 equally likely and mutually exclusive azimuth classes, the probability p of occurrence of azimuths in any class is 1/18.

$s = 507$ total grains counted $\quad p = 1/18$

$q = 1 - p = \frac{17}{18}$ $\quad x = 51$ grains in classes with elongation azimuths within 10° of current direction

$E = sp = \frac{507}{18} = 28 =$ expected number of grains in any 10° class

$$d = x - E = 51 - 28 = 23$$

$$\sigma = (spq)^{\frac{1}{2}} = \left(\frac{507 \cdot 17}{18 \cdot 18}\right)^{\frac{1}{2}} = 5.16$$

$$\delta = \frac{d - \frac{1}{2}}{\sigma} = \frac{23.0 - 0.5}{5.16} = 4.37$$

$$P_\delta = 1 - 2\int_0^{4.37} \phi(t)dt = 1 - 2(0.50000)$$

= 0.00000 = probability of occurrence of a deviation from the expected value, in *any* class, of as much as 23. Hence, the probability of occurrence of this modal value in one of the two classes adjacent to the azimuth of current direction is 2/18·0.00000 = 0.00000.

Comparison of the long-dimension and least-projection elongation distributions gives a slight advantage to the latter. The long-dimension modal frequency of 46 (Fig. 3) is less distinct than the two classes containing 51 least-projection azimuths. In addition, the long-dimension azimuth distribution exhibits a greater standard deviation, a greater skewness, and a mean differing by a larger amount from the true current direction than the corresponding least-projection values. The least-projection method is, therefore, slightly more critical than the long-dimension method—a factor which might prove decisive in an investigation of specimens from environments productive of more nearly homogeneous orientation patterns.

TABLE 1

CHARACTERIZING CONSTANTS OF DISTRIBUTIONS OF AZIMUTHS MEASURED 0°–180° FROM KNOWN CURRENT DIRECTION FOR SAND DEPOSITED BY RUNNING WATER (FLUVIAL DEPOSITIONAL ORIENTATION PATTERN)

	Long-Dimensional Azimuthal Direction (Fig. 3)	Least-Projection Azimuthal Direction (Fig. 4)
Mean	3.3°	0.0°
Standard deviation	48.1°	45.8°
Skewness	−0.136	−0.124
Kurtosis	2.22	2.17

In order to determine the effect of cross-currents on elongation azimuth distributions, the total sample was divided into two portions, the first containing 124 grains from the region in which no cross-currents were observed, the second comprising the 383 grains from the area in which cross-currents were noted (Fig. 4). Table 2 is a summary of the azimuthal distribution of the least-projection tangents. It is apparent that cross-currents have the ef-

fect of increasing the standard deviation, increasing the skewness, and decreasing the kurtosis of the distribution. The fact that the mean azimuth of the portion affected by cross-currents differs from the true current direction by only 0.9° suggests the applicability of this system of analysis to materials from natural environments in which cross-currents are common.

Insufficient data are available from this investigation to determine whether

TABLE 2

CHARACTERIZING CONSTANTS OF DISTRIBUTIONS OF LEAST-PROJECTION AZIMUTHS MEASURED 0°–180° FROM KNOWN CURRENT DIRECTION FOR SAND DEPOSITED BY RUNNING WATER (FLUVIAL DEPOSITIONAL ORIENTATION PATTERN)

	Distribution Where Cross-Currents Are Absent	Distribution Where Cross-Currents Are Present
Mean	177.0°(−3.0°)	0.9°
Standard deviation	39.0°	47.7°
Skewness	0.073	−0.194
Kurtosis	2.54	2.09

the mean or modal azimuth of elongation gives the more accurate portrayal of current direction. The distributions summarized above are so symmetrical that their mean and modal values nearly coincide. It is suggested, however, that the mean is affected by cross-currents to a greater degree than is the mode and, hence, that the mode is the more dependable value.

SUMMARY OF FLUVIAL END-POSITION DATA

The arrangement of grains with their larger ends statistically toward the source of material is demonstrated conclusively by the measurement of area direction. Of the 447 grains amenable to estimation by the visual method, 62 per cent had their greater area toward the current. Since the probability of such a deviation from the expected random value of 50 per cent in a sample of this size is 0.00000,[19] there can be no doubt that the observed preponderance of area direction in the upcurrent position is due to physical factors and does not represent merely chance deviation from a random distribution. The accepted limit of probability of 0.01 as indicative of a preferred distribution is reached by every series of 200 consecutive grains in the sample; hence, area direction data for 200 grains probably could be considered as sufficient to yield reliable results in this type of distribution.[20]

The measurement of end position in the investigation using the method of position of intersection of axes (p. 251) does not indicate the presence of a preferred position of the large ends of the grains. Approximately 52 per cent of the intersections fall on the north side of the midpoint of the long-dimension axis. The probability of obtaining this deviation of 2 per cent from the expected value of 50 per cent, if the distribution were due to chance alone, is 0.467 (nearly 1 in 2). Hence this method is either entirely unsatisfactory or requires measurement of a much larger number of grains than does the method of area direction. Its weakness is attributed to the important effect of slight irregularities of grain outline upon the location of the two axes.

[19] The 0.00000 probability is calculated according to the method illustrated on p. 252.

[20] In order to determine the least number of consecutive grains in the observed distribution required to have a probability of occurrence by chance below the 0.01 significance level, several trial probabilities were calculated. Each of these included data from photographs containing the least excess over the value expected by chance. It was found that the least number of consecutive grains required for a probability of less than 0.01 included 206 grains.

ANALYSIS OF LABORATORY EOLIAN SAND

The problem of determining whether depositional patterns similar to those produced by running water result from wind action was investigated in the laboratory by blowing sand in one direction. In order to produce wind blowing with constant force and limited direction at the source, a current of air from a $\frac{1}{8}$-inch hose was directed downward onto the stream table from a point about $\frac{1}{8}$-inch above the floor of the table. Dry sand from the beach of Lake Michigan, as in the fluvial experiments, was supplied at the air source until a layer of sand had been deposited over the entire table. In the process of deposition many grains moved through the air directly to their place of rest, whereas others moved by jumps similar to the saltation movement of grains in a fluvial environment. A section of the stream table was then removed and the surface of the deposit photographed, following the method previously described. In this manner the projection outlines of 642 grains which had been deposited 2 feet from the sand source were obtained. Measurements of (1) least-projection azimuth and (2) area direction were made on these grains.

SUMMARY OF EOLIAN LEAST-PROJECTION DATA

Data from 642 grains show that for least-projection azimuths a mode of 49 exists in the class adjacent to the wind direction and a secondary mode of nearly equal prominence (48) exists in the 160°–169° class (11°–20° from the current direction) (Fig. 5). The excess of 13.3 ($49.0 - 35.7 = 13.3$) over the expected value of 35.7 has a probability 0.027 of occurring by chance in any class and the probability of 0.003 of chance occurrence in either class adjacent to the current direction. As this value is below the 0.01 significance level, there can be no doubt that the orientation is due to the operation of physical factors.

The characterizing constants obtained by statistical analysis of the eolian least-projection azimuths are shown in Table 3. Here again the mode and mean differ from the current direction by a small amount. Hence, the principles of interpreting current direction from elongation alignment apply to sand from eolian as well as from fluvial environments. The eolian distribution shows a slightly greater standard deviation and smaller kurtosis parameter than the fluvial pattern.

TABLE 3

CHARACTERIZING CONSTANTS OF DISTRIBUTION OF LEAST-PROJECTION AZIMUTHS MEASURED 0°–180° FROM KNOWN CURRENT DIRECTION FOR SAND DEPOSITED BY WIND OF CONSTANT DIRECTION (EOLIAN DEPOSITIONAL ORIENTATION PATTERN)

Mean	1.3°
Standard deviation	49.8°
Skewness	−0.04
Kurtosis	1.94

Although these differences are too small to be analyzed with certainty, they suggest an expectable tendency toward a lesser degree of preferred orientation resulting from wind action than from water currents.

Comparison of the fluvial and eolian azimuth histograms (Figs. 4 and 5) reveals the presence of a secondary mode normal to the current direction in the eolian distribution, a feature indicating that some of the grains rolled about their long axis. If further study reveals that this secondary mode is characteristic of eolian sediments, a means of distinguishing between sands from eolian and fluvial environments is provided. It should be noted, however, that the azimuth distribution may possibly vary more within a single group of eolian or fluvial sedi-

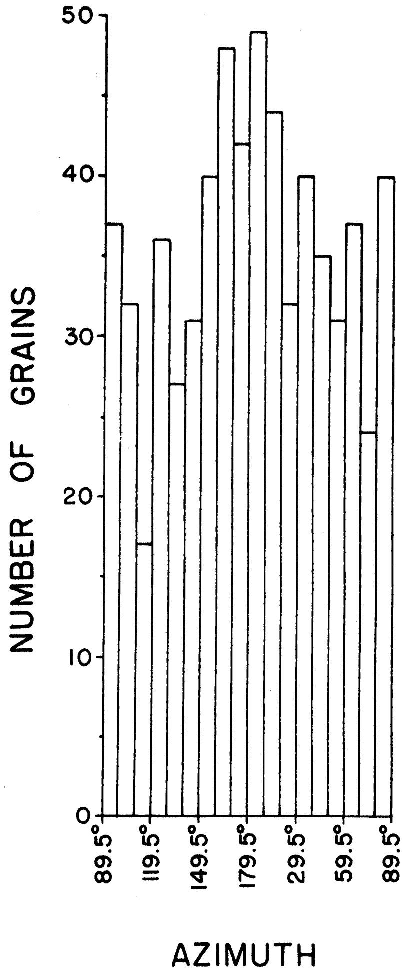


FIG. 5

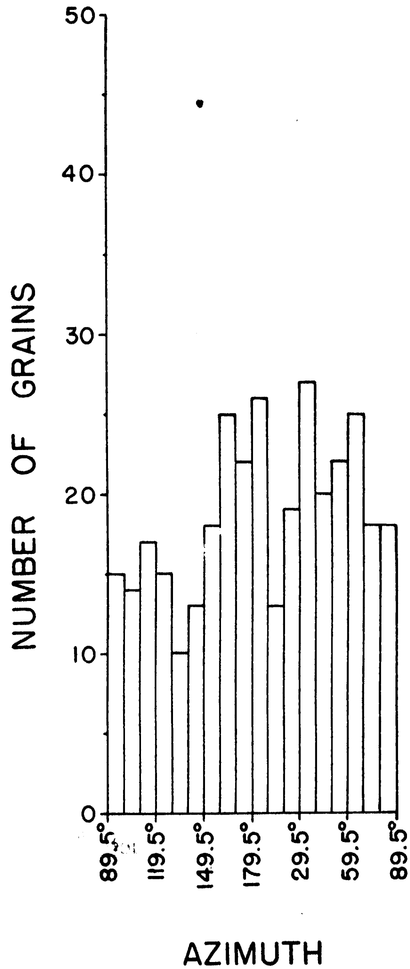


FIG. 6

FIG. 5.—(*Left*) Histogram of least projection azimuth distribution of laboratory eolian sand, showing secondary mode 90° from current direction. Current direction = 0.0° (= 180.0°).

FIG. 6.—(*Right*) Histogram of least projection azimuth distribution of laboratory gravity sand, showing random nature of the distribution. Direction of dip = 0.0° (= 180.0°).

ments, owing to differences in current velocity or other factors not studied, than between these two classes of sediments.

SUMMARY OF EOLIAN AREA DIRECTION DATA

Determination of area direction for the eolian depositional pattern shows that 55 per cent of the grains have their larger ends toward the source. The probability of obtaining this excess over the 50 per cent expected by chance for the 522 grains determined is 0.024. This probability, although greater than the conservative 0.01 limit, is small enough to leave little doubt that the distribution is the effect of the depositing agent and does not represent merely chance deviation from a random distribution.

It is noteworthy that the percentage of grains with area direction toward their source is less than in the fluvial specimens (55 as compared with 62 per cent). Although the actual percentage in each case may be a function of current velocity or some other factor, this difference should be investigated as a possible means of differentiating between sands deposited in fluvial and eolian environments.

ANALYSIS OF LABORATORY GRAVITY SAND

To simulate depositional conditions on the lee side of a dune, dry sand was piled on the sectional stream table and allowed to roll and settle to a position of rest. This was accomplished by inclining the stream table and pouring sand at the top until the angle of repose (32°) was reached and the sand had rolled part way down the trough. Photographs were taken through a binocular microscope set up normal to the surface of the sand, new material being brought into the microscope field by allowing additional sand to roll down from above, until 337 grains had been photographed. The surface of the sand remained at such an angle that grains added at the upper end rolled down either individually or in small landslides.

Measurements were made of (1) least-projection elongation and (2) area direction with respect to the direction of grain movement—that is, the direction of dip of the slope.

The data obtained show that elongation azimuths of grains on a gravity slope are apparently distributed at random (Fig. 6). The mode of elongation direction lies in the class 30°–39° from the direction of dip, and the maximum deviation from the expected value in the grains measured has the probability 0.06 of occurring by chance. Consequently, no interpretation of source direction is possible from such gravity patterns.[21] The lack of preferred elongation is attributed to the small landslides: grains rolling individually down a slope probably tend to have their long axes parallel to the strike of the slope.

Measurements of area direction with respect to the direction of inclination of the slope yield the surprising result that 60 per cent of the grains have their larger ends uphill. The probability of obtaining such a deviation from the expected value of 50 per cent is 0.001. Hence the observed pattern is due to physical causes and does not represent merely chance deviation from a random distribution. It is noteworthy that in this preferred position there is a greater distance between the lower end of the grain and its center of mass than if the larger end of the grain

[21] A method of determining the source direction of material deposited on inclined slopes in dune areas has been described by Parry Reiche, "An Analysis of Cross-lamination: The Coconino Sandstone," *Jour. Geol.*, Vol. XLVI (1938), pp. 905–32.

were downhill. Consequently, more energy is required to roll a grain downhill from this position. Although the absence of preferred grain alignment precludes the application of area-direction measurements to unknown gravity patterns, this energy principle may be of value in future investigations of patterns from other environments.

CONCLUSION

The limited scope of the investigation presented in the preceding pages does not justify the formulation of extensive conclusions concerning dimensional orientation patterns in fine-grained sediments. Nevertheless, it is hoped that certain characteristics which have been observed will serve as a working basis for future study. These observations are summarized as follows:

1. The particles of fine clastic sediments are not oriented at random but exhibit diagnostic patterns which are a product of their depositional environments.

2. A preferred elongation parallel to the current direction exists among grains in fluvial and eolian environments; long-dimension orientation in gravity sands is apparently random.

3. In fluvial, eolian, and gravity environments the larger ends of the grains, as determined by area-direction measurements, tend to lie toward the source.

4. The source direction of fluvial and eolian sediments can be determined accurately by combining measurements of least-projection azimuth and area direction.

5. Differences observed among specimens from the three environments suggest that the agent of deposition of a sediment may be determined from certain characteristics of its dimensional orientation pattern.

ACKNOWLEDGMENTS.—The authors wish to express appreciation to Professor Mason E. Wescott, of the Department of Mathematics, Northwestern University, for his suggestions and constructive criticism of the statistical methods employed herein.

The investigation was conducted in the laboratories of the Department of Geology and Geography, Northwestern University, and this paper, to a large extent, represents the Master's thesis of the junior author.

THE STONES RIVER EQUIVALENTS IN APPALACHIAN REGION[1]

BYRON N. COOPER
Virginia Geological Survey, Charlottesville

ABSTRACT

The commonly accepted correlation of the Stones River group of central Tennessee with the Chazy of New York and with beds immediately succeeding the Beekmantown or Knox dolomite in the Appalachian region is no longer tenable. Ulrich's recognition of two Chazyan groups in the Appalachian Valley—the Stones River and overlying Blount—was based upon mistaken identification of formations.

In southwestern Virginia the fauna of the Murfreesboro limestone (basal Stones River) occurs in the Peery limestone about 475 feet above the Beekmantown. The Peery is directly overlain by the Benbolt limestone, which correlates faunally with the Pierce and Ridley formations of the Central Basin of Tennessee. In Bland County, Virginia, Murfreesboro-Ridley equivalents overlie the Athens formation of Ulrich's Blount group. The Lebanon of central Tennessee is linked faunally with the Witten limestone, which corresponds to the beds identified by Butts as Lowville in the Clinch Mountain belt of the Middle Ordovician in Virginia and Tennessee.

Both the stratigraphic position of Stones River equivalents in the Appalachian region and the character of Stones River faunas indicate strongly that the entire Stones River group is post-Chazyan.

ORIGIN AND USAGE OF THE NAME "STONES RIVER"

The name "Stones River group" was originally proposed by J. M. Safford[2] for a succession of limestones exposed along Stone(s) River in the Central Basin of Tennessee. Three subdivisions were named and described: (1) the Stones River beds, at the base, about 75 feet thick; (2) the lower Lebanon limestone; and (3) the upper Lebanon limestone, with a thickness of 110–30 feet. In his next description[3] of this succession the name "Stones River" was replaced by "Trenton or Lebanon formation," and the lithology and fossils of the beds were described for the first time. Five subdivisions were named, in ascending order: (1) Central limestone, about 100 feet thick, forming the "bottom-rock" of the Central Basin and corresponding to the lowest of the three divisions described in 1851; (2) Pierce limestone, composed of thin flaggy beds, with a total thickness of 27 feet; (3) Ridley limestone, thick-bedded and aggregating 95 feet; (4) Glade limestone, light-blue flaggy limestone about 120 feet thick, forming the Cedar Glades of central Tennessee; and (5) the Carters Creek limestone, thick-bedded, dove-gray, totaling 50–100 feet. These units are well displayed in Rutherford County, Tennessee, in the middle of the Central Basin.

In eastern Tennessee, Safford[4] referred to the succession above the Knox and below the Clinch sandstone as the "Trenton and Nashville series," in which he recognized seven divisions. All were given informal lithologic names, except the lowest, which was called the "Maclurea Limestone." Whether Safford then regarded the Trenton group of the Central Basin to be well represented in eastern Tennessee is not clear; but, since the lowest division—the Maclurea limestone—was assigned to the "Trenton and Nashville series," it is only reasonable to infer that Safford regarded the Maclurea limestone to be equivalent to

[1] Published by permission of the state geologist of Virginia.

[2] "The Silurian Basin of Middle Tennessee, with Notices of the Strata Surrounding It," *Amer. Jour. Sci.*, Vol. XII (2d ser., 1851), pp. 352–61.

[3] Safford, *The Geology of Tennessee* (Nashville, 1869), pp. 258–68.

[4] *Ibid.*, p. 228.

part of the "Trenton (Lebanon) formation" of central Tennessee. Safford, according to N. H. Winchell and E. O. Ulrich,[5] abandoned the name "Stones River" in 1869, under the impression that the group was strictly equivalent to the New York Trenton.

J. M. Safford and J. B. Killebrew[6] subsequently referred to the succession as the "Lebanon (Trenton) group." The name "Lenoir" was introduced to replace †Maclurea limestone. It was defined as overlying the Knox and underlying the variegated marbles of the Knoxville region, which were believed to represent the "Lebanon group" of the Central Basin. The Lenoir was described as the correlative of the New York Chazy. Thus Safford and Killebrew in 1876 regarded the entire "Lebanon group" as younger than the Chazy.

Winchell and Ulrich[7] later revived the name "Stones River group," upon concluding that the Central, Pierce, Ridley, and Glade limestones were equivalent to the Birdseye limestone of New York rather than to the New York Trenton, as Safford and Killebrew had believed. So close did Winchell and Ulrich regard the correlation of the Birdseye and Stones River beds that they then considered it likely that the name "Stones River" would eventually replace the nongeographic name "Birdseye." At that time Winchell and Ulrich considered the Birdseye older than the Black River, following the usage of Black River inaugurated by James Hall[8] in 1847.

[5] *The Geology of Minnesota*, Vol. III (Part II, 1897), p. xc.

[6] *The Elementary Geology of Tennessee* (Nashville, 1876), pp. 130–33.

[7] P. xc of ftn. 5 (1897).

[8] "Description of the Organic Remains of the Lower Division of the New York System," *Paleon. N.Y.*, Vol. I (1847), pp. 14, 37, 46.

In 1900 Safford and Killebrew[9] accepted the revived name "Stones River" and proposed formal geographic names for the Central and Glade limestones—Murfreesboro and Lebanon, respectively. The Carters Creek limestone was taken out of the Stones River group, the name shortened to "Carter," and the formation referred to the overlying Nashville group. In a table of formations[10] the Stones River group was classed with the Chazy and placed in a position to indicate equivalency of the group to the Lenoir limestone. Apparently this is the first record of the now time-honored correlation of the Chazy and Stones River.

Three years later, C. W. Hayes and Ulrich[11] put the Carter limestone back in the Stones River group and inaugurated the spelling "Carters."

Ulrich and Schuchert [12] expressed the opinion that most of the Stones River was pre-Birdseye. They considered the lower Stones River ". . . . about equivalent in time to the Chazy and Levis." Also, they maintained that the "extreme top of the Stones River is equivalent to the Lowville limestone."

Further changes in Ulrich's interpretation of the Stones River group appeared in 1908.[13] Cushing, in receipt of two letters from Ulrich—one dated November 19, 1907, and the other March 25, 1908—stated Ulrich's views as follows:

[9] *The Elements of the Geology of Tennessee* (Nashville, 1900), pp. 125–26.

[10] Safford and Killebrew, pp. 104–5 of ftn. 9 (1900).

[11] "Description of the Columbia Quadrangle, Tennessee," *U.S. Geol. Surv. Folio 95* (1903), correlation chart.

[12] "Paleozoic Seas and Barriers in Eastern North America," *N.Y. State Mus. Bull. 52* (1902), p. 641.

[13] H. P. Cushing, "Lower Portion of the Paleozoic Section of Northeastern New York," *Bull. Geol. Soc. Amer.*, Vol. XIX (1908), pp. 155–76.

> Ulrich no longer regards the Lowville as of upper Stones River age, but puts it above that formation. . . . . It should be stated also that the lower and middle divisions of the Stones River formation are regarded as essentially equivalent to lower and middle Chazy in time, but that the comparatively small Chazy basin was at this time entirely separated from the much larger sea to the south and west in which the Stones River deposits were laid down. "Finally the upper Stones River or Pamelia is only partially represented in the Chazy section by dove reef limestone at the base of the upper Chazy." [Last sentence quoted by Cushing from Ulrich's letter of March 25, 1908.]

Ulrich[14] reiterated his conclusion that the Pamelia, Crown Point, and Day Point limestones of New York were equivalent in time to the upper, middle, and lower divisions of the Stones River group of central Tennessee and that the Stones River group is therefore entirely Chazyan. He continued to regard the Carters limestone as the topmost formation of the Stones River group and made the Chazyan an Ordovician series composed of the Stones River and overlying Blount groups. Ulrich's Stones River group of eastern Tennessee was made up of the Mosheim (lower Stones River), Lenoir limestone (middle Stones River) and the upper, unnamed division, which was said to be developed along the western margin of the Appalachian Valley. By classifying the Mosheim and Lenoir limestones with the Stones River, Ulrich inaugurated the usage of the name "Stones River" for a widespread stratigraphic division of the Ordovician system. Thereupon the name also became a time term for the interval during which the Mosheim, Murfreesboro, Pierce, Ridley, Lenoir, Lebanon, and Carters formations were deposited.

[14] Revision of the Paleozoic Systems," *Bull. Geol. Soc. Amer.*, Vol. XXII (1911), p. 640, Pl. 27.

R. S. Bassler[15] continued to regard the Mosheim as the basal formation of the Stones River. He took the Carters limestone out of the Stones River and made it the basal division of the Black River group. This change was adhered to by J. J. Galloway,[16] who also considered the Stones River to be Chazyan.

Butts[17] interpreted the Mosheim as older than the Stones River group of central Tennessee, in which he included only the Murfreesboro, Pierce, Ridley, and Lebanon. The most significant statement made by Butts about the Stones River concerns the supposed equivalency of the Lenoir and Ridley. He said:

> They [*Maclurea magna*] also serve to correlate the Lenoir unmistakably with the middle part of the Chazy limestone of the type locality in northeastern New York. *Maclurea magna* occurs rarely in the Ridley limestone of the Stones River group in middle Tennessee, and by this means the Lenoir is correlated with the Ridley.[18]

Bassler[19] had previously correlated the Lenoir with the Pierce, but as late as 1911 Ulrich[20] refrained from correlating the Lenoir with any single formation in the Central Basin. He pointed out that the Lenoir and the Stones River beds were deposited in separate basins, with characteristic faunas which were not suitable for precise correlation.

[15] "Bibliographic Index of American Ordovician and Silurian Fossils," *U.S. Nat. Mus. Bull. 92* (Vol. II, 1915), Pl. 1.

[16] "Geology and Natural Resources of Rutherford County, Tennessee," *Tenn. Geol. Surv. Bull. 22* (1919), pp. 30–31, 32–45.

[17] "The Paleozoic Rocks" in "Geology of Alabama," *Ala. Geol. Surv. Spec. Rept. 14* (1926), opp. p. 80.

[18] *Ibid.*, p. 105.

[19] Pl. 1 of ftn. 15 (1915).

[20] P. 539 of ftn. 14 (1911).

In 1928 Ulrich[21] and Butts concluded from a study of the section exposed along Yellow Branch, Lee County, Virginia, that the Mosheim was younger than the Murfreesboro. With this change, Ulrich's[22] revised Chazy classification was as follows:

- Chazyan series
  - Little Oak limestone (Alabama)
  - Blount group
    - Ottosee shale and limestone
    - Tellico sandstone
    - Athens shale
    - Whitesburg limestone
    - Holston limestone
  - Stones River group
    - Lebanon limestone
    - Ridley-Lenoir limestones
    - Mosheim limestone
    - Murfreesboro limestone

This is the classification used by most geologists during the last fifteen years.

In 1939 Ulrich[23] abandoned his previous interpretation of the Chazyan and Stones River and introduced a new classification, shown in Figure 1. The disposition of the type Stones River formations is noteworthy. The Lebanon was placed in the Black River; and the Pierce-Ridley, long considered older than any part of Ulrich's Blount group, was placed above the Ottosee. Neither Blount nor Stones River appears in the classification. Two new names—Strasburg and Speers Ferry—were introduced, though without description or definition. This postulated succession negates Ulrich's prior contention regarding the existence of two Chazyan groups—the Stones River and overlying Blount. He continued to classify the Ridley and Pierce with the Chazy and maintained that the Murfreesboro is older than the Mosheim and Lenoir of the Appalachian Valley.

Although Ulrich, Bassler,[24] Butts,[25] and many others have held that the Stones River is Chazy, P. E. Raymond[26] has maintained since 1905 that the Stones River is post-Chazy. He stated:

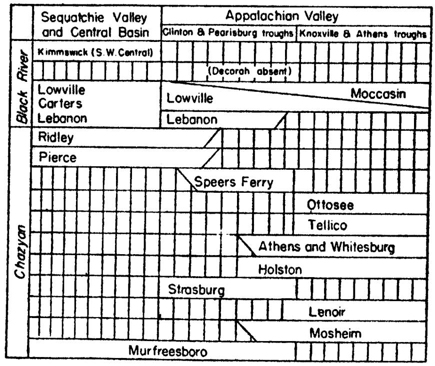


FIG. 1.—Correlation of Middle Ordovician formations in Tennessee (Ulrich, 1939).

Comparing the large percentage of forms common to the Stones River and to the Black River and Trenton with the low percentage—less than 5 per cent—of forms common to the Chazy and Trenton, it becomes evident that the Stones River and Trenton are faunally much more closely connected than are the Chazy and Trenton. These close relationships of the fauna of the Stones River to that of the Trenton, coupled with the stratigraphy, suggest

[21] "Ordovician Trilobites of the Family Telephidae and Concerned Stratigraphic Correlations." *Proc. U.S. Nat. Mus.*, Vol. LXXVI, Art. 21 (1929), p. 78.

[22] *Ibid.*, p. 73.

[23] "The Murfreesboro Limestone in Missouri and Arkansas and Some Related Facts and Probabilities," *Kan. Geol. Soc. 15th Ann. Field Conf. Guidebook* (1939), p. 106.

[24] "Stratigraphy of the Central Basin of Tennessee," *Tenn. Div. Geol. Bull. 38* (1932), pp. 48–58.

[25] Butts *et al.*, "The Southern Appalachian Region," *Guidebook 3, 16th Internat. Geol. Cong.* (1932), pp. 13–14; Butts, "Geologic Map of the Appalachian Valley of Virginia with Explanatory Text," *Va. Geol. Surv. Bull. 42* (1933), pp. 13–14, and "Geology of the Appalachian Valley in Virginia," *Va. Geol. Surv. Bull. 52* (Part I, 1940), pp. 119–78.

[26] "The Fauna of the Chazy Limestone," *Amer. Jour. Sci.*, Vol. XX (4th ser., 1905), pp. 353–82.

that the whole Stones River is younger than the Chazy.[27]

Schuchert[28] also believed that the Stones River is post-Chazy and stated:

> The writer made a list of 89 Stones River species, other than bryozoans. . . . . Of these 89 species, 7 are said to begin in the Chazy . . . . , and about 57 occur either in the Lebanon (a formation accepted by nearly every stratigrapher as of Black River age), or in other accepted Black River strata.

USAGE OF "STONES RIVER" IN THE APPALACHIAN REGION

The term "Stones River" has long been used in the Appalachian region for various successions of strata, generally including a fossiliferous zone characterized by *Maclurites magnus* Lesueur. In Pennsylvania, G. W. Stose[29] and Ulrich assigned 1,050 feet of beds in one section and at Martinsburg, West Virginia, 675 feet to the Stones River formation. A fauna collected from the middle, *Maclurites*-bearing division of his Stones River formation was interpreted by him to represent the middle Chazy of New York. He said:

> The fossils and the lithologic character of the formation are so nearly the same as those of the Stones River limestone of middle Tennessee that they are regarded as identical. Furthermore, the formation has been recognized by Ulrich and others at intervals along the Appalachian Valley southward into Tennessee. The name Stones River is therefore applied to this series of pure, fine-grained limestones.[30]

In Maryland, R. S. Bassler[31] assigned 1,100 feet in a generalized section to the Stones River. The middle fossiliferous division, as in Pennsylvania, contains *Maclurites magnus* and is about 200 feet thick. This part of the so-called Stones River is correlated by Bassler with the Crown Point limestone (middle Chazy) of New York.

In northern Virginia and particularly in the vicinity of Strasburg, Shenandoah County, Bassler[32] recognized 900 feet of beds in the Stones River; but, as Butts[33] has shown, most of this thickness is Beekmantown dolomite and limestone.

G. P. Grimsley[34] recognized a threefold division of the Stones River in the eastern Panhandle of West Virginia and gave the names "Lower, Middle, and Upper Stones River Limestone" to the three divisions previously recognized, but not named, by Stose in Pennsylvania.

D. B. Reger[35] has used "Stones River (Chickamauga) limestone" for the entire succession of limestones above the Beekmantown and below the Moccasin limestone (Fig. 2). According to Reger, the Stones River limestone in Mercer and Monroe counties contains "no vestige" of the lower Stones River (Murfreesboro).

In Virginia, Butts[36] has used the name "Stones River" for a succession ranging in thickness from a few feet to more than 1,300 feet, which directly overlies the Beekmantown and extends up to the top of the supposed Lenoir. The lowest division, occurring mainly south of New River and northwest of the Saltville

[27] *Ibid.*, p. 368.

[28] *Stratigraphy of the Eastern and Central United States* (New York: John Wiley & Sons, 1943), pp. 474–75.

[29] "Description of the Mercersburg-Chambersburg District," *U.S. Geol. Surv. Folio 170* (1909), pp. 7–8.

[30] *Ibid.*, p. 6.

[31] "Cambrian and Ordovician," *Md. Geol. Surv. Rept.* (1919), p. 120.

[32] "The Cement Resources of Virginia West of the Blue Ridge," *Va. Geol. Surv. Bull. II-A* (1909), p. 78.

[33] Pp. 197–98 of ftn. 25 (1940).

[34] "Jefferson, Berkeley, and Morgan Counties," *W.Va. Geol. Surv. Rept.* (1916), pp. 265–71.

[35] "Mercer, Monroe, and Summers Counties," *W.Va. Geol. Surv. Rept., 1926*, pp. 620–30.

[36] Pp. 119–47 of ftn. 25 (1940).

fault, was referred to as the "Blackford facies of the Murfreesboro formation." It contains characteristically reddish dolomitic siltstones with included pebbles of chert and dolomite, light-gray mealy shales, and fossiliferous cherty layers. In some places, particularly at St. Clair Station, Tazewell County, Lenoir was correlated with the Ridley of central Tennessee. Butts correlated the Murfreesboro with the basal member of the Chazy of Champlain Valley in New York and the Lenoir with the Crown Point limestone.

In summary, "Stones River" as a formation and group name has been used

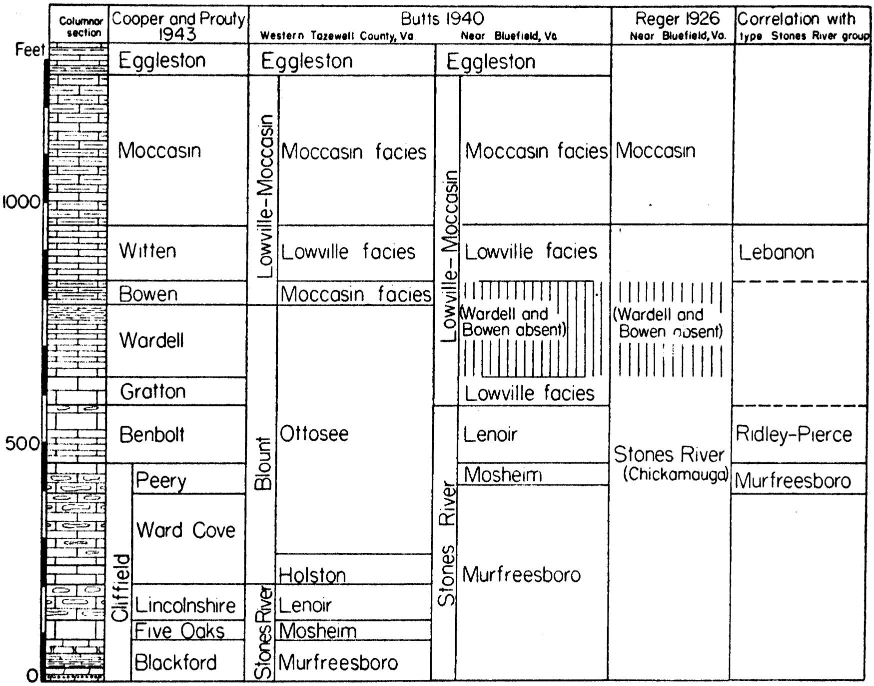


FIG. 2.—Position of Stones River equivalents in belts northwest of Clinch Mountain in southwestern Virginia.

Virginia—less than 4 miles from Mercer County, West Virginia, wherein Reger[37] believed the Murfreesboro to be absent—Butts recognizes a locally developed limestone facies of the Murfreesboro, with a thickness of more than 1,000 feet. Beds which he believed to represent the Lenoir and Mosheim were also classified[38] with the Stones River, and the widely in the Appalachian region, but the various usages are inconsistent. In spite of the many similarly emphatic correlations of beds in the Appalachian region with the Stones River group, very little faunal evidence has been presented to substantiate the correlations. Until recently,[39] Murfreesboro, wherever used

[37] P. 622 of ftn. 35 (1926).

[38] Butts, p. 147 of ftn. 25 (1940).

[39] B. N. Cooper and C. E. Prouty, "Stratigraphy of the Lower Middle Ordovician of Tazewell County, Virginia," *Bull. Geol. Soc. Amer.*, Vol. LIV (1943), pp. 819–86; B. N. Cooper, "Geology and Mineral

in the Appalachian region, has been generally regarded as the oldest post-Beekmantown formation. The Carters is no longer considered part of the Stones River. Various classifications of the Stones River formations are shown in Table 1.

STRATIGRAPHIC POSITION OF THE MURFREESBORO IN CENTRAL TENNESSEE

In the original description of the formation[40] the Murfreesboro was described as the oldest limestone in the Central Basin and as consisting of the lowest 70 feet of exposed beds. Ulrich[41] regarded the type Murfreesboro as probably resting on the Beekmantown or Knox. The true position of the type Murfreesboro did not become generally known until 1940, when K. E. Born[42] published a detailed description of the cuttings obtained from the Basin Oil and Gas Company's Henry Harrell No. 1 well, which was drilled in the circular outcrop area of the Murfreesboro at the type locality, Rutherford County, Tennessee. He identified 380 feet of limestone above the Knox and below the type Murfreesboro. Thus Ulrich's supposition that the Murfreesboro of the type locality is probably just above the Beekmantown is erroneous. The true position of the type Murfreesboro with respect to the subsurface Knox dolomite is of special interest because of the remarkably similar position of the Murfreesboro fauna in the Ordovician succession of southwestern Virginia.

IN THE APPALACHIAN REGION

Midway between Wittens Mills and Five Oaks, Tazewell County, Virginia, the Peery member of the Cliffield formation[43]—400–475 feet above the Beekmantown dolomite—contains an assemblage of fossils found in the type Murfreesboro (Fig. 2). Among the species identified from the cherty beds of the Peery member are *Lophospira procera* Ulrich, *L. perangulata* (Hall), *Helicotoma* aff. *H. tennesseensis* Ulrich and Scofield, *H. declivis* Ulrich, *Trochonema bellulum* Ulrich, *Ophiletina sublaxa depressa* Ulrich and Scofield, *Plectoceras bondi* (Safford), *Tetradium syringoporoides* Ulrich, *Leperditia fabulites* (Conrad), *L. pinguis* Butts, *Phragmolites* sp., and peculiar forms resembling closely *Polylopia billingsi* (Safford). In the same section Butts recognizes the Murfreesboro, Mosheim, and Lenoir (Fig. 2), but the top of his Lenoir is at least 200 feet stratigraphically below beds containing the Murfreesboro fossils mentioned above. Also in this section, the beds comprising the upper part of the Peery limestone member are a typical calcilutyte,[44] like the Mosheim. At two other localities in Tazewell County—one at Pounding Mill[45] and the other at St. Clair—this calcilutyte has been identified by Butts[46] as Mosheim. That this zone of Mosheim lithology is several hundred feet above the limestone identified by Butts as Mosheim in the section near Wittens Mills has been shown previously.[47] It can also be positively affirmed that the cherty limestones with Murfreesboro

Resources of the Burkes Garden Quadrangle, Virginia," *Va. Geol. Surv. Bull. 60* (1944), p. 71.

[40] Safford and Killebrew, p. 125 of ftn. 9 (1900).

[41] Pl. 27 of ftn. 14 (1911).

[42] "Lower Ordovician Sand Zones ('St. Peter') in Middle Tennessee," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXIV (1940), pp. 1641–62.

[43] Cooper and Prouty, pp. 866–68 of ftn. 39 (1943).

[44] *Ibid.*, p. 826.

[45] Butts *et al.*, p. 65 of ftn. 25 (1932).

[46] Pp. 123–24 of ftn. 25 (1940).

[47] Cooper and Prouty, pp. 852–53 of ftn. 39 (1943).

TABLE 1

CLASSIFICATIONS OF THE STONES RIVER GROUP

<table>
<tr><th colspan="2">Safford, 1851</th><th colspan="2">Safford, 1869.</th><th colspan="2">Safford and Killebrew, 1876</th><th colspan="2">Ulrich and Winchell, 1897</th></tr>
<tr><td rowspan="5">Stones River group</td><td rowspan="2">Upper Lebanon limestone</td><td rowspan="5">Trenton (Lebanon) formation</td><td>Carter's Creek limestone</td><td colspan="2">(Eastern Tenn.) Nashville group</td><td rowspan="5">Stones River group</td><td>Carter's Creek limestone</td></tr>
<tr><td>Glade limestone</td><td rowspan="3">Lebanon group</td><td rowspan="3">380 feet of marble in the Knoxville area</td><td>Glade limestone</td></tr>
<tr><td rowspan="2">Lower Lebanon limestone</td><td>Ridley limestone</td><td>Ridley limestone</td></tr>
<tr><td>Pierce limestone</td><td>Pierce limestone</td></tr>
<tr><td>Stones River beds</td><td>Central limestone</td><td colspan="2">Lenoir limestone</td><td>Central limestone</td></tr>
</table>

<table>
<tr><th colspan="2">Safford and Killebrew, 1900</th><th colspan="2">Ulrich and Hayes, 1903</th><th colspan="2">Ulrich, 1911 (Generalized Section)</th><th colspan="2">Ulrich, 1911 (Eastern Tenn.)</th></tr>
<tr><td colspan="2">Carter limestone</td><td rowspan="5">Stones River group</td><td>Carters limestone</td><td rowspan="6">Stones River group</td><td>Carters limestone</td><td colspan="2">Lowville limestone</td></tr>
<tr><td rowspan="4">Stones River group</td><td>Lebanon limestone</td><td>Lebanon limestone</td><td>Lebanon limestone</td><td rowspan="5">Stones River group</td><td rowspan="2">Upper Stones River</td></tr>
<tr><td>Ridley limestone</td><td>Ridley limestone</td><td>Ridley limestone</td></tr>
<tr><td>Pierce limestone</td><td>Pierce limestone</td><td>Pierce limestone</td><td rowspan="2">Lenoir limestone</td></tr>
<tr><td>Murfreesboro limestone</td><td>Murfreesboro limestone</td><td>Murfreesboro limestone</td></tr>
<tr><td></td><td></td><td></td><td></td><td>Mosheim limestone</td><td>Mosheim limestone</td></tr>
</table>

<table>
<tr><th colspan="2">Bassler, 1915</th><th colspan="2">Galloway, 1919</th><th colspan="2">Butts, 1926</th><th colspan="2">Ulrich, 1929</th></tr>
<tr><td colspan="2">Carters limestone</td><td colspan="2">Carters limestone</td><td colspan="2">Carters limestone</td><td colspan="2">Lowville limestone</td></tr>
<tr><td rowspan="5">Stones River group</td><td>Lebanon limestone</td><td rowspan="4">Stones River group</td><td>Lebanon limestone</td><td rowspan="4">Stones River group</td><td>Lebanon limestone</td><td rowspan="5">Stones River group</td><td>Lebanon limestone</td></tr>
<tr><td>Ridley limestone</td><td>Ridley limestone</td><td>Ridley limestone</td><td>Ridley limestone</td></tr>
<tr><td>Pierce limestone</td><td>Pierce limestone</td><td>Pierce limestone</td><td>Pierce limestone</td></tr>
<tr><td>Murfreesboro limestone</td><td>Murfreesboro limestone</td><td>Murfreesboro limestone</td><td>Mosheim limestone</td></tr>
<tr><td>Mosheim limestone</td><td></td><td></td><td colspan="2">Mosheim limestone</td><td>Murfreesboro limestone</td></tr>
</table>

TABLE 1—*Continued*

Bassler, 1932		Butts, 1940 (Appalachian Valley)		Butts, 1940 (Central Tenn.)		Ulrich, 1939 (Central Tenn.)	
Carters limestone		Blount group		Carters limestone		Black River group	Lowville limestone
Stones River group	Lebanon limestone	Stones River group	Lenoir limestone	Stones River group	Lebanon limestone	Black River group	Carters limestone
Stones River group		Stones River group		Stones River group		Black River group	Lebanon limestone
Stones River group	Ridley limestone	Stones River group	Mosheim limestone	Stones River group	Ridley limestone	Chazyan	Ridley limestone
Stones River group	Pierce limestone	Stones River group		Stones River group	Pierce limestone	Chazyan	Pierce limestone
Stones River group		Stones River group		Stones River group	*Hiatus?*	Chazyan	*Major hiatus*
Stones River group	Murfreesboro limestone	Stones River group	Murfreesboro limestone	Stones River group	Murfreesboro limestone	Chazyan	Murfreesboro limestone

fossils are at least 200 feet above the *Sowerbyites triseptatus–Dinorthis atavoides* zone, which, with few exceptions, constitutes Butts's Lenoir limestone in Tazewell, Russell, and Scott counties of southwestern Virginia.

Along Yellow Branch, Lee County, Virginia, Butts[48] reported Murfreesboro fossils from a thin zone 175 feet above the Beekmantown. This zone is part of a 100-foot succession between two almost equally thick zones of calcilutyte. The upper calcilutyte was identified by Butts and Ulrich as the Mosheim, and this identification is the basis for their claim that the Mosheim is younger than the Murfreesboro. Since two similar zones of calcilutyte have been confused with the Mosheim in some other localities in southwestern Virginia, there is justification for doubting the identity of the Mosheim as determined by Butts and Ulrich in the Yellow Branch section. Butts's Mosheim of the Yellow Branch section occupies the same stratigraphic position as the calcilutyte at the top of the Peery limestone member, being above beds containing the Murfreesboro fauna and directly below argillaceous layers containing *Öpikina*, *Mimella*, and bryozoans such as occur in the Pierce and Ridley of central Tennessee. This calcilutyte near Wittens Mills is part of Butts's Ottosee. In view of the confusion regarding the identity and position of the Mosheim in southwestern Virginia, it can no longer be regarded as a certainty that the Mosheim is younger than the Murfreesboro.

One mile north of Ceres, Bland County, Virginia, the Peery limestone overlies a succession of black limestones and shales containing the *Nemagraptus gracilis* fauna characteristic of a formation identified by Butts as Athens (Fig. 3). The Murfreesboro, which is the equivalent of the Peery, is therefore younger than most of Ulrich's Blount formations. According to Ulrich, the Murfreesboro belongs at the base of the limestone section—hundreds of feet below the Athens!

It is difficult to establish precisely the position of the Murfreesboro with respect to the type Mosheim and Lenoir, since all three have not been positively identified in any section. The *Sowerbyites triseptatus–Dinorthis atavoides* zone—200-

[48] Pp. 120–21 of ftn. 25 (1940).

250 feet below the true Murfreesboro equivalent in the section near Wittens Mills—is Ulrich's "Strasburg (upper Lenoir)" and Butts's Lenoir in the also occurs in limestone intercalations in dolomitic chert-breccias of Butts's "Blackford facies of the Murfreesboro formation" along Moccasin Creek, south

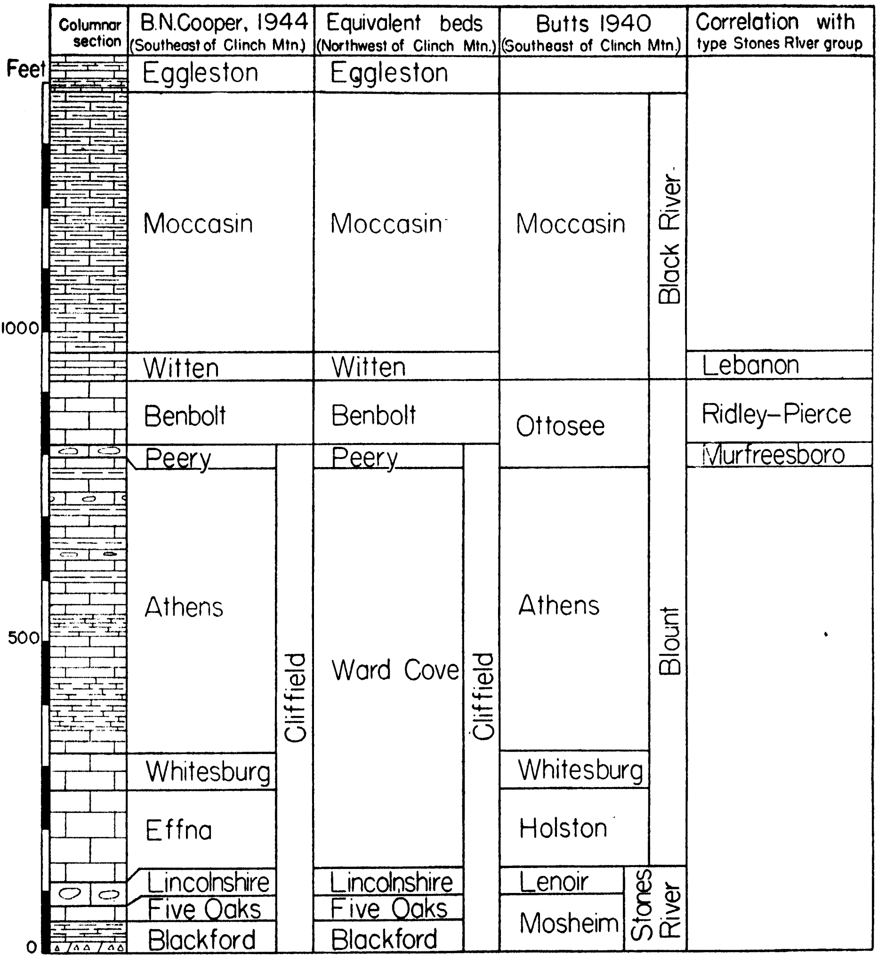


FIG. 3.—Position of Stones River equivalents in belts southeast of Clinch Mountain in southwestern Virginia.

Clinch Mountain belt of Virginia and Tennessee. Whether this zone is a part of the type Lenoir is unknown; probably the typical Lenoir is older. The lowest beds of the type Lenoir are crowded with *Rostricellula pristina* (Raymond), which of Tumbez, Russell County, Virginia. A normal development of the *Dinorthis atavoides–Sowerbyites triseptatus* zone begins about 175 feet higher in the Tumbez section. If the *Sowerbyites-Dinorthis* zone is part of the type Lenoir, most, if

not all, of the succession down to the Beekmantown in southwestern Virginia could be correlated with the Lenoir. The Murfreesboro equivalent is definitely placed considerably above the *Sowerbyites-Dinorthis* zone near Wittens Mills and in Bland County succeeds the Athens formation of Butts, which is supposed to be considerably younger than the type Lenoir.

STRATIGRAPHIC POSITION OF THE PIERCE-RIDLEY IN THE APPALACHIANS

The Benbolt limestone, which succeeds the Peery limestone in southwestern Virginia, contains many fossils found in the Pierce and Ridley of the Central Basin. Perhaps the most apparent and convincing similarity is the abundance of strophomenoid brachiopods in the three formations. *Öpikina*, particularly *Ö.* cf. *Ö. minnesotensis*, and *Strophomena* are prominent. *Camerella plicata* Schuchert and Cooper and similar forms of *Campylorthis*, *Glyptorthis*, *Paurorthis*, and *Hesperorthis* also occur in the three formations. "*Protorhyncha ridleyana*," recently redefined and renamed *Ancistrorhyncha costata* Ulrich and Cooper and usually regarded as a Ridley guide fossil, was collected by G. A. Cooper from beds in Rye Cove, Scott County, Virginia, which are referred by the writer to the Benbolt and classed by Butts with the Ottosee. The same, or closely allied, forms of the bryozoan genera *Anolotichia*, *Chasmatopora*, *Graptodictya*, *Nicholsonella*, and *Rhinidictya* occur in the Benbolt and in either the Pierce or Ridley. The position of the Benbolt, directly above the Murfreesboro equivalent in Tazewell County, Virginia, is corroborative evidence for the identification of the Benbolt as the equivalent of the Pierce and Ridley. *Fascifera subcarinata* and *Stromatocerium rugosum*, which occur in the Ridley, are also found in the lower part of the Wardell formation (upper Ottosee) in southwestern Virginia and along U.S. Route 25E, about $1\frac{1}{2}$ miles southeast of Evans Ferry, Grainger County, Tennessee. Thus the Benbolt probably does not represent all the Ridley.

STRATIGRAPHIC POSITION OF THE LEBANON

The equivalent of the Lebanon limestone of central Tennessee is the Witten limestone,[49] which corresponds to Butts's Lowville along Clinch Mountain in southwestern Virginia and Tennessee. Not only is the general lithology of the Lebanon and Witten much the same, but many fossils are common to the two. Among these are *Cryptophragmus antiquatus* Raymond, *Pionodema minuscula* Willard, similar forms of *Sowerbyella*, *Rostricellula*, *Doleroides*, and *Zygospira*,[50] and abundance of bryozoans of the genera *Escharopora*, *Orbignyella*, *Rhinidictya*, and *Nicholsonella*. Also similar, if not identical, types of fucoids and *Camarocladia*-type sponges are found in both the Lebanon and Witten.

In southwestern Virginia the Benbolt and Witten are separated by 200–400 feet of beds, comprising the Gratton, Wardell, and Bowen formations (Fig. 2). Possibly part of this succession may be represented in the upper Ridley and lower Lebanon of central Tennessee; but, judging from the greater thickness and the greater diversity of faunas, the Benbolt-Witten succession in southwestern

[49] Cooper and Prouty, pp. 877–79 of ftn. 39 (1943).

[50] Information on the occurrence of *Cryptophragmus antiquatus*, *Pionodema minuscula*, and other fossils in the Lebanon was furnished by G. Arthur Cooper, who, in company with Josiah Bridge, collected them from the middle part of the Lebanon limestone in central Tennessee.

Virginia would seem to represent a more nearly complete record of deposition than the Pierce, Ridley, and Lebanon of central Tennessee. A hiatus between the Ridley and Lebanon, therefore, seems probable.

AGE OF THE STONES RIVER BEDS

Ulrich's conception of the Chazyan series in the Appalachian region, embracing the Stones River and Blount groups, is untenable. The Murfreesboro is considerably younger than the Mosheim and Lenoir, and it is above the *Nemagraptus gracilis* zone that occurs in the Athens of Virginia and the Normanskill of New York. Most geologists have considered the typical Lenoir to be equivalent to the Crown Point limestone of the middle Chazy of New York. P. E. Raymond[51] concurs in this correlation and assigns the Normanskill-Athens graptolite zone to the upper Chazy. Although Ulrich places the Ottosee in the Chazy, no faunal basis for the correlation was ever given, and none is known to exist. Although it is very doubtful whether beds as young as the Athens are actual correlatives of the upper division of the Chazy, this correlation could be accepted without regarding any of the Stones River as Chazyan. Now that the Ridley and Lenoir are known not to be equivalent, the main "basis" for linking the Stones River with the Chazy is removed.

The fossils in the Stones River formations and in their Appalachian equivalents strongly suggest a post-Chazyan age. The main elements in the brachiopod faunas, particularly species of *Öpikina*, *Campylorthis*, *Fascifera*, *Hesperorthis*, *Doleroides*, *Zygospira*, and *Strophomena*, are characteristic of formations generally conceded to be Black River in the Mississippi Valley region. Other Stones River fossils considered to be post-Chazy include *Stromatocerium rugosum*, *Foerstephyllum halli*, *Fletcheria*, and *Lambeophyllum*—all of which occur in the Benbolt-Witten and Pierce-Lebanon successions. Even *Tetradium syringoporoides*, long regarded as an infallible index of the Chazy, is reported to be common in the lower Black River of New York and Ontario.[52] *Maclurites magnus*, which is widely considered as an index of the Chazy, has been reported from the Murfreesboro, Pierce, Ridley, and Lebanon of central Tennessee[53] and from beds identified by Butts as Ottosee.[54] Either this species has long range or it has been loosely identified. A few other species, notably *Leperditia fabulites* and *Lophospira perangulata*, which occur in the Stones River, are reported from the Chazy; but they have been cited also from the Pamelia and Lowville. As previously indicated by Raymond and Schuchert, the dominance in the Stones River of forms essentially foreign to the Chazy and characteristic of post-Chazy formations favors a post-Chazy age for the entire Stones River group of central Tennessee.

Cryptophragmus antiquatus, which is common in the Pamelia and lower Lowville of New York, occurs in association with other Black River fossils in the Witten and Lebanon limestones. If the Lebanon and equivalent beds in the Appalachian region are Black River and if

[51] "Correlation of the Ordovician Strata of the Baltic Basin with Those of Eastern North America," *Bull. Harvard Mus. Comp. Zoöl.*, Vol. LVI, No. 3 (1916), pp. 236–37.

[52] V. J. Okulitch, "Some Black River Corals," *Trans. Royal Soc. Canada*, Vol. XXXII (3d ser., 1938), pp. 93–94.

[53] Bassler "Stratigraphy of the Central Basin of Tennessee," *Tenn. Div. Geol. Bull. 38* (1932), pp. 51–58.

[54] Cooper and Prouty, pp. 868–71 of ftn. 39 (1943).

the entire Stones River group is post-Chazy, a new lower Middle Ordovician group should be recognized. Although the existence of such a division is strongly indicated by what is already known of the fossils and stratigraphy of the lower Middle Ordovician in southwestern Virginia, the precise lower limit is still in question and must remain so until the upper limit of the Chazy in the Appalachian region is accurately determined. Although New York is certain to remain the "hub" of Ordovician correlations, the succession in that area must not be adjudged necessarily "ideal" for the whole United States.

Since the stratigraphic relations of the type Murfreesboro to unexposed older beds is obscure and since an integral, if not the most characteristic part, of the Stones River group—the Lebanon limestone—seems unmistakably linked with the Black River, there is no reason for continued usage of the name "Stones River group." If the Ridley, Pierce, and Murfreesboro are to be recognized as a part of a new post-Chazyan, pre-Black River group, the name "Stones River" should not be revived. The new group should be named and defined from an Appalachian section, where relations with subjacent and overlying formations are fully apparent.

SUMMARY OF CONCLUSIONS

1. The Murfreesboro, long considered the oldest post-Beekmantown formation in the Appalachian region, is considerably younger than the Mosheim and Lenoir. In Bland County, Virginia, the Peery limestone, which is the faunal equivalent of the Murfreesboro, overlies a formation of black limestones and shales containing the *Nemagraptus gracilis* fauna which is characteristic of Butts's Athens in southwestern Virginia.

2. The Benbolt limestone, which immediately succeeds the Peery limestone in Bland and Tazewell counties, Virginia, correlates with the Pierce and Ridley of central Tennessee.

3. The Witten limestone of southwestern Virginia, constituting a part of what Butts has called "Lowville" limestone, is faunally linked with the Lebanon of the Central Basin.

4. Since the Benbolt and Witten are separated by three formations, two of them with rather distinctive faunas, the succession of southwestern Virginia would seem to represent a fuller record of deposition than the Pierce-Lebanon of the Central Basin. An unconformity probably exists between the Lebanon and Ridley in the latter area.

5. Confusion regarding the relative ages of the Murfreesboro, Mosheim, and Lenoir resulted from failure to recognize the occurrence of more than one zone of "characteristic" Mosheim lithology. Two such zones have been called "Mosheim" in southwestern Virginia. It seems to be clearly demonstrated that the bed identified as "Mosheim" along Yellow Branch, Lee County, is far above the position of the type Mosheim.

6. Since the Murfreesboro is younger than the Lenoir, the Ridley and Lenoir cannot be equivalent.

7. Ulrich's concept of a "Chazyan series," embracing the Stones River and overlying Blount groups, is invalid. The Mosheim and Lenoir are older than Stones River, and the Stones River formations are younger than the greater part of Ulrich's Blount group.

8. Fossils purported to link the Stones River with the New York Chazy are either forms known to have long range or ones which probably have been loosely identified. They have little index value.

9. The predominance of post-Chazyan

forms in the Stones River beds and in equivalent formations in the Appalachian region would strongly suggest that the entire Stones River group is younger than the Chazy. The uppermost Stones River formation, the Lebanon, seems clearly to represent part of the Black River.

10. The available evidence seems to indicate the probable existence of a post-Chazy, pre-Black River group in the Appalachian region—a group which includes beds at least as old as the Peery limestone and possibly as young as the Bowen or the Wardell formations. The Murfreesboro, Pierce, and Ridley would represent a substantial part, though probably not all, of this group. The precise limits of the division have not yet been determined, for lack of sufficient information about the fossils and about the position of the upper Chazyan boundary in the Appalachian section.

11. "Stones River" as a general time-rock term should no longer be used, nor should the name be revived if a post-Chazyan, pre-Black River group is recognized.

ORIGIN OF GRANITE DOMES IN THE SOUTH-EASTERN PIEDMONT

WILLIAM A. WHITE
University of North Carolina, Chapel Hill

ABSTRACT

The forms of the granite domes of the Southeast, which heretofore have been attributed to exfoliation, are regarded as the product of granular disintegration brought about by chemical weathering.

INTRODUCTION

There are many dome-shaped exposures of granite throughout the Southeastern Piedmont. The largest, the well-known Stone Mountain of De Kalb County, Georgia, is well known; but its many smaller counterparts throughout the Piedmont of Georgia and the Carolinas are very little known. The domes vary greatly in size from the massive Stone Mountain itself, which is about a mile and a half long and some 650 feet higher than the surrounding peneplain, down to small inconspicuous bosses, little larger than residual boulders. They are all alike, however, in having smooth, spheroidal surfaces without sharp protuberances or re-entrants. Most are composed of granite, but some few are gneissic. None show visible joint systems.

Heretofore they have been explained as the product of exfoliation; but the writer, after several years of field work throughout the region in which they occur, has come to believe that they have resulted largely from the action of other processes. The validity of the exfoliation theory of origin for granite domes in certain other climatic regions is not questioned, but those which have developed from the granites and gneisses of the southeast do not show enough evidence of exfoliation to justify the assumption that it has played a dominant or even an important part in their formation. It is true that some of the southeastern domes show a little exfoliation; but in most instances the surface is smooth and unbroken, and evidence of exfoliation is rarely seen (Fig. 1).

The exfoliation theory of dome development stands upon the essentially sound reasoning that an unjointed mass of homogeneous rock will be attacked most readily on those parts of its surface which have the smallest radii of curvature and that the continuation of such selective attack will reduce the mass to a spheroidal form. But this principle applies equally well to the denudation caused by any other weathering agent which is nondirectional in its attack. In the southeastern states the climate is warm and humid, and the remarkable development of chemical weathering has long been a matter of comment. Hydration, oxidation, and carbonation are nondirectional in their attack and should be adequate to reduce an irregular-shaped mass to spheroidal form provided the rock resistance is uniform. Since the granites and gneisses from which the domes have been sculptured are quite homogeneous and without joints, there is little reason to believe that their resistance to chemical weathering should be differential.

While chemical weathering has usual-

ly been most pronounced in areas underlain by jointed or schistose rocks, there is good evidence that it has also had a significant effect upon the unjointed masses of granite and gneiss from which these domes have been shaped. Quarrymen working on them usually find it necessary to remove 6 inches to several feet of sap before they encounter unaltered rock. These are not impressive thicknesses, but kaolinization character-

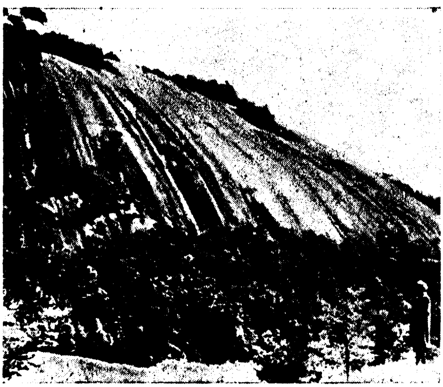


FIG. 1.—Stone Mountain in Wilkes County, North Carolina.

istically takes place on the intergranular surfaces of the feldspars, and the rock at the surface of the exposure tends to break up into a gruss which is washed off the steep bare slopes as quickly as it is formed. Many observers have noted this intergranular alteration. L. E. Smith[1] has made petrographic studies showing it in the surficial phases of the unjointed granite masses of the South Carolina Piedmont, and the writer has made similar studies in North Carolina.

EVIDENCE FOR EXFOLIATION

It is evident that exfoliation and granular disintegration are coexistent as sculpturing agents on the domes, and their relative importance should be determined by a comparison of the evidence for each. Considering first the case for exfoliation, if a dome were dominantly the result of its action, one would expect to find that fact manifest in two ways. (1) Since few exfoliation spalls detach themselves from the parent-mass in the form of complete lenses, one would expect to find the surface of the dome covered by truncated remnants of spalls which had partially fallen away. There would probably be overlapping of such remnants, and the surface of the dome would have a somewhat imbricate appearance. Half Dome in Yosemite (Fig. 2) offers an excellent example of such a surface. (2) If the surface of the dome had been produced by exfoliation, at its base one would expect to find a talus slope composed of fallen and broken spalls.

In the case of the domes of the Southeast neither of these criteria is satisfied. As stated above, there is very little evidence of exfoliation on the dome surfaces. Looking at the face of Stone Mountain in Wilkes County, North Carolina, the observer can see the broken edges of no more than two or three spalls from any one viewpoint. And at the base of the mountain only a few remnants of fallen spalls can be found. This observation applies equally well to the attendant bosses and minor domes which appear near by. On Stone Mountain in De Kalb County, Georgia, there is somewhat better evidence of exfoliation; and it seems to have been a slightly more important factor in the denudation of that mass. Even there, however, the broken edges of spalls are so rounded by normal weathering (granular disintegration) that they are not conspicuous. Other less well-known domes throughout the region

[1] "Weather Pits in Granite of the Southern Piedmont," *Jour. Geomorph.*, Vol. II (1941), p. 125.

show the effects of exfoliation in varying intensity, but most of them are as little affected by it as is Stone Mountain in Wilkes County, North Carolina. Some are affected even less.

It is true that there seems to be a latent tendency toward hypogene exfoliation in all the domes, but there does not appear to be much evidence that it has ever been activated by wholly natural agencies. In every place where it can be recognized, it has been initiated artificially by the rapid removal of overburden in the process of quarrying. Most of the natural spalls are quite thin and seem to be the result of other causes.

It is possible that the domes were produced by exfoliation under different climatic conditions in a previous geologic age. However, this possibility seems remote, for there is no evidence of an arid climate in this region later than Triassic time.

In brief, if a dome were sculptured by exfoliation, the evidence for the action of that agency should be spectacularly displayed as remnants of broken spalls both adhering to the dome surface and composing a talus slope at its base.

FIG. 2.—Half Dome in Yosemite Valley, California, after F. E. Matthes. Photograph by courtesy of the U.S. Geological Survey.

EVIDENCE FOR GRANULAR DISINTEGRATION

On the other hand, if a dome had developed through granular disintegration, the evidence to prove it should be somewhat obscure. Formation of gruss is a grain-by-grain process. As soon as a grain has been loosened from the parent-mass, it is washed off the steeply sloping surface either to become part of the bed load of the small drainage ways at the

base of the dome or to be incorporated in alluvial fans surrounding it. Because of the slow rate at which the individual grains are released from the parent-mass, there is small tendency for fans to develop; but in every location which is favorable to the detention of sediment there are deposits of the coarse debris produced by granular disintegration. Furthermore, every dome of the writer's acquaintance is surrounded for significant distances by the young soils which result from granular disintegration. The dominant series is the Louisburg, a light-colored sandy soil without definite profile development. Field evidence[2] suggests that it develops largely upon unjointed granites and gneisses as the product of granular disintegration. The intimacy of its association with the domes is exemplified by the soils map shown in Figure 3. This has been reproduced from unit-area maps of the Soil Conservation Service, United States Department of Agriculture,[3] and

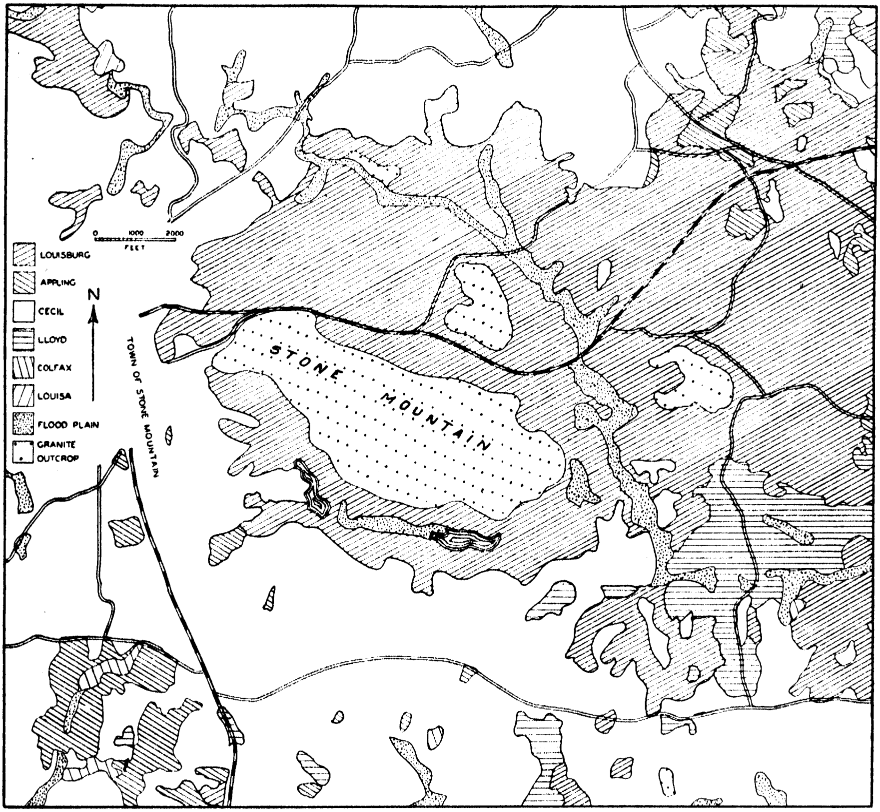


FIG. 3.—Map of area around Stone Mountain in De Kalb County, Georgia, showing distribution of soil series.

[2] W. A. White, "Determining Factors in the Coloration of Granite Soils in the Southeastern Piedmont," *Amer. Jour. Sci.*, Vol. CCXLII, No. 7 (1944), pp. 361–63.

[3] P. H. Montgomery, "Erosion and Related Land Use Conditions of the Lloyd Shoals Reservoir Watershed, Georgia," Phys. Surv. Div., Soil Cons.

tension to the original veneer will have similar protective qualities and will tend to preserve the sharp boundary between the dome and the surrounding soil. This process may possibly explain the sharp knick points which characteristically appear at the bases of domes, and (although the writer does not like to extend his conclusions to regions unfamiliar to him) it may have some connection with the genesis of the bornhardts of East Africa.

In conclusion, the writer would like to suggest that perhaps geologists have erroneously considered granite domes to be unusual land forms, produced only by very special conditions. It would probably be more catholic to regard them as the expected form wherever nonjointed homogeneous rocks are subjected to the attack of any nondirectional agency of denudation. The particular agency would be determined by the local climate, but the resulting land forms should be essentially similar.

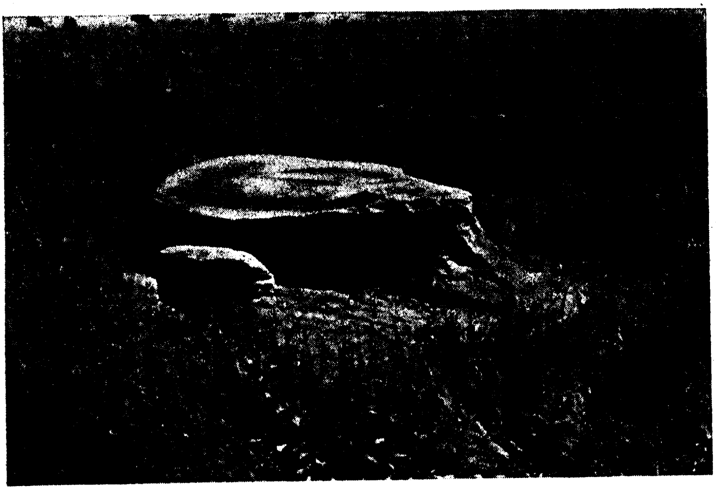


FIG. 4.—Granite outcrop protected by indurated veneer, 2 miles southwest of Wendell in eastern Wake County, North Carolina.

REVIEWS

The Pleistocene Geology of Iowa: Special Report. By GEORGE E. KAY AND OTHERS. Part I: "The Pre-Illinoian Pleistocene Geology of Iowa," by GEORGE F. KAY and EARL T. APFEL. First published in "Iowa Geol. Survey," Vol. XXXIV (1928). Pp. 304; figs. 63; pls. 3; Pl. I in pocket. Part II: "The Illinoian and Post-Illinoian Pleistocene Geology of Iowa," by GEORGE F. KAY and JACK B. GRAHAM. First published in "Iowa Geol. Survey," Vol. XXXVIII (1943). Pp. 262; figs. 89. Part III: "The Bibliography of the Pleistocene Geology of Iowa," by GEORGE F. KAY. [1943]. Iowa City, Iowa: Iowa Geological Survey, [1943]. Pp. 55.

This volume is labeled *Special Report* because it combines parts of two volumes of the Iowa Geological Survey, to which is added "The Bibliography of the Pleistocene Geology of Iowa." As a special volume it constitutes a monographic treatment of the Pleistocene geology of Iowa.

In the Introduction to Part II, "The Illinoian and Post-Illinoian Pleistocene Geology of Iowa," the authors call attention to the fact that certain advances had been made since the publication of the report on "The Pre-Illinoian Pleistocene Geology of Iowa" in 1928—especially with respect to the classification of the Pleistocene—and they set forth the classifications shown in Tables 1 and 2 as those which are now recognized by the Iowa Geological Survey, one for the (upper) Mississippi Valley and the other for the state of Iowa.

If the reader bears in mind that fifteen years intervened between the publication of Parts I and II and that in the meantime revisions of the classification were made by Kay and Leighton, the principal discrepancies between Parts I and II will be clarified. The Iowan glacial stage had become a part of the Wisconsin, and the Peorian interglacial stage had become an *intra*glacial substage in the Iowa section. Furthermore, Kay had chosen to recognize the Nebraskan (glacial) and Aftonian (interglacial) as constituting an epoch to which he applied the name "Grandian," the Kansan (glacial) and Yarmouth (interglacial) another epoch which he called "Ottumwan," the Illinoian (glacial) and Sangamon (interglacial) a third epoch named "Centralian," and the Wisconsin (glacial) and Recent a fourth epoch labeled "Eldoran." Accordingly, the chapter headings of Part II are different in form from the corresponding chapter headings of Part I.

The reviewer has followed closely for thirty years the work which Kay has pursued so vigorously on the Pleistocene geology of Iowa. In 1911, upon the death of his predecessor, the late Professor Samuel Calvin, who had been state geologist almost continuously since the organization of the present Survey, Kay began his Pleistocene studies in Iowa while the writer was a student at the University of Iowa. Later, when the writer was assisting Dr. William C. Alden in a review of the evidences of the Iowan drift in northeastern Iowa, Kay invited them to participate in a field conference in southern Iowa to observe his findings with reference to the nature and origin of the "gumbo" which lay at the surface of the Kansan drift. Still later, after the writer of this review had become state geologist of Illinois, frequent interstate conferences were held over a period of twenty years.

Not a single field season was missed by Kay. He carried his investigations in a systematic manner from area to area and from problem to problem until he had covered the entire state and the whole Pleistocene section, including the bedrock topography upon which the glacial deposits rest. During each college year he dedicated himself to the other exacting responsibilities of professor, head of the department of geology, dean of the College of Liberal Arts, director of the State Geological Survey, and scientific scholar and writer. Nevertheless the monograph now under our review portrays a patient documentation of the features and phenomena observed, a critical and fair-minded review of the work and views of his predecessors and contemporaries, an astute analysis of the problems, a clear presentation of his own interpretations and concepts, and a frank recognition of baffling problems. At some points in the report the reviewer finds himself desiring to observe the field evidence in the light of his own

experience or from his own point of view, but nowhere does he question the devotion of the investigator to his problem.

That Iowa contains a rich record of the Pleistocene period and that Kay expended great energy in unraveling it will impress the readers of this outstanding volume. By his work the fundamental stratigraphy of the Pleistocene has been made secure. Hence this complete report has a special place in geological literature and is warmly received.

TABLE 1

CLASSIFICATION OF PLEISTOCENE GEOLOGY IN THE MISSISSIPPI VALLEY

Period (System)	Epoch (Series)	Age (Stage)	Subage (Substage)
Pleistocene or Glacial	Eldoran	Recent Wisconsin	Mankato Cary Tazewell Iowan
	Centralian	Sangamon Illinoian	
	Ottumwan	Yarmouth Kansan	
	Grandian	Aftonian Nebraskan	

The discussion of each of the glacial drifts and interglacial phenomena is meticulous from both field and laboratory standpoints. In the case of each glacial drift, the criteria that may be used in its discrimination are set forth, followed by a discussion of its distribution, origin, changes it underwent subsequently, and typical sections in various areas and in different stratigraphic relationships. Then there follows a thoroughgoing description of the subdivisions of the whole profile of weathering, including the unweathered parent-material.

An unusual feature of the report is the presentation of laboratory results of sedimentation studies from the physical and chemical standpoints. The tills, silts, sands, and gravels and their weathered zones were sampled and subjected to careful laboratory testing and analysis by the junior authors and advanced students. This phase of study included their clastic texture, size-grade distribution, lithology, roundness of pebbles, and chemical composition. The resultant data were then assembled, compared, and contrasted to provide valuable information on the original constitution of the glacial-laid, water-laid, and wind-laid materials, on the changes they underwent by weathering in different zones of the weathering profile, and on the value of various criteria.

Considerable attention was given by Cornelia Cameron to the molluscan faunas of the Loveland loess and the Peorian loess, and by George H. Lane to the climatic and floral interpretation of peat zones, based on pollen grains. Further work of the foregoing character will contribute valuably to our knowledge and understanding of the sediments of the Pleistocene and to the conditions which prevailed. The mammalian and other vertebrate record contained in this report was derived mainly from the literature, but Kay makes a valuable contribution in pointing out that the stratigraphic position of a considerable number of the sources is open to serious question and that this field of knowledge should be thoroughly reviewed and modernized to provide a proper interpretation of the prevailing climatic and ecologic conditions.

Part I is devoted to separate chapters on the bedrock surface, topography and drainage, a review of investigations and classification of the Pleistocene deposits, the Nebraskan glacial stage, the Aftonian interglacial stage, the Kansan glacial stage, and the Yarmouth interglacial stage, together with an admirable summary under the heading of "Concluding Statements."

TABLE 2

CLASSIFICATION OF THE PLEISTOCENE GEOLOGY OF IOWA

Period (System)	Epoch (Series)	Age (Stage)	Subage (Substage)
Pleistocene or Glacial	Eldoran	Recent Wisconsin	Mankato Peorian Iowan
	Centralian	Sangamon Illinoian	
	Ottumwan	Yarmouth Kansan	
	Grandian	Aftonian Nebraskan	

The viewpoint is expressed that the preglacial topography of Iowa is by no means the present topography of the bedrock surface, either beneath the drift or in the so-called "Driftless Area" of northeastern Iowa. Kay

and Apfel point out that the present relief of northeastern Iowa is 200–400 feet greater than the relief of the bedrock surface under the glacial drift. This greater relief is thought to have been developed following the derangement of drainage by the Nebraskan glacier, which diverted the master-stream to the course of the present Mississippi River. In this way deeper dissection was initiated and accomplished in northeastern Iowa during the Aftonian interval. This concept of deep Aftonian erosion in northeastern Iowa is that of Trowbridge, who found that the Nebraskan drift there consists of scattered patches whose lower limit is about that of the Lancaster erosion surface. The authors also point out that the plotting of the average of the highest elevations of the bedrock in all parts of the state reveals that a topographic low—a reduced drainage basin—extended across Iowa from north to south, entering the state a little east of the middle of the north line and leaving it in the southeast corner.

From this knowledge they derive the concept that, at the close of the Tertiary, Iowa possessed a subdued topography near or at the Lancaster erosion level and that the incised valleys of northeastern Iowa and the steep-walled, drift-buried valleys in the Kansan drift area and some elsewhere were carved chiefly during the Aftonian interglacial stage. These concepts will await testing when sufficient well data and other information are in hand.

With respect to the present topography and drainage of Iowa, Kay and Apfel furnish a clear general picture of the variety of topographic conditions prevailing in the state. Among the points of outstanding interest are the Pleistocene age of the deep dissection of the so-called "Driftless Area" in northeastern Iowa; the mature stage of erosion of most of southern Iowa, involving successive grade levels, and more advanced erosion from east to west toward the Missouri River; the disappearance of tabular divides on the Kansan drift of northwestern Iowa from south to north; the drift-mantled erosional topography of the Iowan drift area in eastern Iowa and its border of loess-mantled erosional topography; the unique loess depositional topography of western Iowa bordering the Missouri flood plain, changing eastward into a belt of loess-mantled erosional topography; and the youthful glacial features of the drift-depositional topography of the Des Moines lobe.

Chapter iii of Part I gives a thorough and instructive review of the history of investigations of the Pleistocene deposits of Iowa, interwoven with the development of the classifications. C. A. White, in 1870, first recognized the glacial origin of the surficial materials in Iowa, but McGee made the first important contributions to the unraveling of the glacial history of the deposits in 1878. Chamberlin had already differentiated in a masterly manner two drifts in Wisconsin and established the record of an interglacial epoch, as had Winchell in Minnesota.

Kay and Apfel review the comprehensive work of McGee in a spirit of understanding of the state of knowledge of that time and without criticism of his failure to recognize three tills instead of two. They emphasize, instead, his full treatment, for that time, of the Upper Till and the Lower Till of northeastern Iowa and its associated loess, subsequently named "East Iowan" and then "Iowan," and "Kansan," respectively; his recognition of the youth of the Upper Till compared with the gumbo-surfaced drift (present Kansan) of southern Iowa and its greater age than the Des Moines lobe (present Wisconsin). Had McGee recognized the presence of three tills, both Chamberlin and McGee might have viewed the Afton Junction section in a different light when they visited it in 1893, made proper correlations with the tills of northeastern Iowa, and saved later controversy regarding nomenclature.

These miscorrelations were discovered by Bain and recognized by Chamberlin in a field conference in 1896, and because the name "Kansan" had in the meantime been applied to the uppermost drift in areas bordering and lying outside the Iowan drift, Chamberlin acceded somewhat hesitantly to the designation of the upper till at the Afton Junction section as Kansan instead of the lower till. Then in his editorial that autumn in the *Journal of Geology* he applied the name "Albertan" (after Dawson) to the lower till. It seems difficult to understand, from this distance, how the upper till at Afton Junction, with loess overlying it unconformably, could have been correlated with McGee's Upper Till (Iowan), inasmuch as McGee had regarded the Upper Till and the loess as of the same age.

The organization of the Iowa Geological Survey in 1892 came at an opportune time for Calvin, Norton, Bain, Udden, Shimek, Tilton, Beyer, and others to pursue vigorously Pleistocene and other studies during the rest of the

1890's and the early 1900's, when knowledge of the Pleistocene was rapidly developing. By 1896, Leverett, of the United States Geological Survey, had discovered the Illinoian stage of glaciation in Illinois and southeastern Iowa, and Shimek had set forth his eolian theory of origin for the loess deposits. By 1898 the names of the principal glacial and interglacial stages had been virtually agreed upon for the Pleistocene of the upper Mississippi Valley by the leading students, with the exception of the Nebraskan, which was suggested by Shimek in 1909. Thus the youthful Iowa Geological Survey shared in this very constructive scientific period.

The controversy that ensued later with regard to the Iowan drift is thoroughly described and analyzed by the authors without prejudice. As a result of this controversy and the additional critical studies made, the Iowan drift and its relationships came to be more fully known and appreciated for its special features.

The first contributions of Kay in regard to the nature and origin of the gumbo on the tabular divides of the Kansan drift of southern Iowa began to appear in 1916; and, as the reader follows the history of his and Apfel's investigations of the gumbotil of the Nebraskan, Kansan, and Illinoian drifts, he cannot help realizing that this later period of refined and detailed studies marks another constructive period for Pleistocene geology. The Aftonian, Yarmouth, and Sangamon interglacial horizons in Iowa became as firmly established as any other part of our whole geologic column, and the use of the Kansan and Nebraskan gumbotil horizons made possible the areal mapping of the Nebraskan drift over a large part of southwestern Iowa and the general contouring of the original surfaces of the Nebraskan and Kansan drift over a large part of the state. Perhaps less notable but yet important was Kay's work on the Loveland formation and the Loveland loess, the latter of late Sangamon age. The position of the Loveland loess above the Illinoian drift and below the Iowan drift is one of the conclusive criteria for fixing the age difference of these two drift sheets and for fixing the upper limit of the Sangamon interglacial horizon.

It will pay the young student of the Pleistocene to read thoroughly this chapter on the history of investigations and of the classifications of the Pleistocene of Iowa, because, in addition to setting forth the subjects mentioned above, it covers many other important features, including the work on the Iowan drift of northwestern Iowa by Carman, the fossil-bearing gravels of southwestern Iowa by Calvin and Shimek, the Pleistocene mammals of Iowa by Hay, the history of the Des Moines River Valley by Lees, the history of Lake Calvin by Schoewe, and several other special studies.

Illuminated by the authors' "History of the Investigations and Classifications of the Pleistocene Deposits of Iowa," the chapters that follow constitute a thoroughgoing treatise of present knowledge of the stratigraphy and history of the Nebraskan glacial stage, the Aftonian interglacial stage, the Kansan glacial stage, and the Yarmouth interglacial stage.

These chapters are methodically organized. The Nebraskan and Kansan drifts are described and discussed with respect to their stratigraphic relations, distribution, origin, changes, typical sections, drift phases (profiles of weathering), and thickness.

The chapter on the Aftonian discusses the record of that interval, sections representative of it, the Nebraskan gumbotil, erosion during the interval, Aftonian loess, and the life of that epoch.

The chapter on the Yarmouth includes a statement of the record; a discussion of the Buchanan interval (post-Kansan, pre-Iowan); descriptions of the weathered products of the Kansan drift beneath the Illinoian drift and outside the Illinoian area; the buried soils and vegetal materials; erosion; silts and sands; the Loveland formation; and the record of life in the Yarmouth.

The authors point out that the Nebraskan drift in many places "can be identified with certainty only if one can establish definitely its relations to certain interglacial materials the age of which is known." The Nebraskan ice, from the Keewatin center, covered all of Iowa; and when it melted, it left drift of sufficient thickness and so distributed over the bedrock surface as to produce a plain with minor surface irregularities. The probable average thickness of the drift for the state is estimated at more than 100 feet. It is difficult to distinguish it from overlying Kansan lithologically except in northwestern Iowa, where it contains a large amount of Cretaceous shale material. The Kansan ice, also from the Keewatin center, covered the state, with the exception of the so-called "Driftless Area," and deposited a massive sheet of drift over the gumbo-surfaced and eroded Nebraskan, leaving a strikingly level or nearly level ground-moraine plain.

It is emphasized that the gumbotils are the most satisfactory horizon markers in the state. It is believed that, when each was formed, the ground-moraine plains were low lying and that dissection of these plains was subsequently initiated by uplift. The reviewer, without denying that there may have been uplift, is of the opinion, after making an extensive study of the Illinoian drift plain in Illinois, that the assumption of uplift is not necessary because the essential topographic condition for the formation of gumbotil is a degree of flatness essential to poor surface and subsurface drainage. Gumbotil is found on the Illinoian drift on plains lying at different levels, but plains sufficiently extensive to have poor drainage.

The authors note from their studies that the so-called "Aftonian gravels" are not of Aftonian age but of Nebraskan, some of which were at the surface and became weathered before the Kansan ice invasion. They suggest that they be called "weathered Nebraskan gravels" and that the name "Aftonian gravels" be dropped. Likewise the sands and gravels of western Iowa are contemporaneous in age with the tills with which they are associated.

The numerous outcrops of gumbotil under the Kansan drift in southwestern Iowa indicate that extensive valleys there of Aftonian age were few. Elsewhere erosion was great, which is especially clear in northeastern Iowa.

Thin loess, referred to the Aftonian, is present in Lyon and Union counties, and thin "loess-like clay" in Shelby County.

As noted hitherto, Part II was prepared several years subsequent to Part I, and Kay's junior associate was Jack B. Graham. This part bears the mark of Kay's new classification in which he combines the Nebraskan and Aftonian to make the Grandian epoch, the Kansan and Yarmouth the Ottumwan epoch, the Illinoian and Sangamon the Centralian epoch, and the Wisconsin and Recent the Eldoran epoch, all of them comprising the Pleistocene period. R. T. Chamberlin has taken exception to ranking the Pleistocene as a system or period for reasons which he set forth with impressive clarity in a recent number of the *Journal of Geology*.

As for the pairing of a glacial stage and the immediately following interglacial stage to form an epoch, the reviewer believes that there is a degree of justification for this in that the altered and secondary materials of the weathered zone of a drift sheet, where overlain by the next younger drift, can then be assigned to the epoch to which they belong, being partly of the glacial age and partly of the interglacial age.

In accordance with this new grouping the chapter headings of Part II run as follows: chapter i, "The Centralian Epoch (Series): The Illinoian Glacial Age (Stage)"; chapter ii, "The Centralian Epoch (Series): The Sangamon Interglacial Age (Stage)"; chapter iii, "The Eldoran Epoch (Series): The Wisconsin Glacial Age (Stage)"; chapter iv, "The Eldoran Epoch (Series): The Recent Interglacial Age (Stage)." The treatment of the glacial drifts and interglacial deposits is, on the whole, similar to that followed in Part I.

Part II also introduces the feature of profiles of weathering, which subject had its development in this country mostly subsequent to the publication of Part I. In this period Leighton and MacClintock amplified and extended the concept of the gumbotil profile to include the "silttil profile," developed under conditions of good drainage in rolling topography, and the mesotil profile, developed under conditions intermediate between good drainage and poor drainage. They also introduced the concept of second-cycle profiles resulting from headward erosion changing poor drainage conditions to good drainage and permitting the colloids of gumbotil to be carried out to form the product silttil (second-cycle).

On page 13 of Part II, Kay and Graham present two new classifications, one for the Mississippi Valley and the other for Iowa. The latter differs from the former in that two of the Wisconsin substages—Tazewell and Cary—are not represented in Iowa by glacial deposits but by Peorian loess. Hence the name "Peorian" is substituted in the Iowa section for Tazewell and Cary. In this connection it is well to draw attention to the fact that the Peorian loess exposed beneath the eastern edge of the Mankato drift of Iowa has a shallow zone of leaching, thus recording that the Peorian substage here is represented both by deposition of loess and by weathering. It is also well to remember that the loess along the immediate eastern and southern borders of the Iowan drift is mostly late Iowan in age, having been blown from the exposed fringe of bare-surfaced Iowan drift while the Iowan ice was still melting away. Along the Mississippi and Missouri rivers there may also be loess of Tazewell, Cary, and Mankato ages as well as Iowan, which had their source from valley-train silts of those times.

The Pleistocene student will find special in-

terest in the discussion of the Loveland formation and its significance, in the well-rounded data concerning the Iowan drift, and in the full treatment of the Peorian loess, including mechanical, mineral, and chemical analyses and the paleontology and ecology of the molluskan fauna and flora. Greater emphasis should have been placed, the reviewer believes, upon the original content and nature of the Peorian loess deposits. The authors cite certain sections of Peorian loess exposed beneath Mankato drift at the southern edge of the Des Moines lobe as "unusual" because there is much woody material within the body of the loess. Such sections are unusual in the sense that they are comparatively few, but they have special significance. The reviewer has found that woody material is common in sections of Peorian loess in Illinois where the loess is buried and preserved from weathering by the overlying Wisconsin drift. The prevailing conception of loess comes from the widespread exposures of loess that has been oxidized and, through the oxidation, has lost its original woody content. A proper conception of the original nature of the Peorian loess of the upper Mississippi Valley is to be derived from a study of those sections which have been preserved from weathering by an overlying cover of younger Wisconsin drift.

Part III, "The Bibliography of the Pleistocene Geology of Iowa," was prepared by Professor Kay with the assistance of several graduate students. This excellent addition to this monograph is a valuable source of reference not only for students of the Pleistocene of Iowa but for those working in bordering states or in special phases of the Glacial period of North America.

It is fortunate, indeed, that the junior authors of Parts I and II were privileged to be associated with the senior author in the preparation of this special report on *The Pleistocene Geology of Iowa*. Such a relationship as this which Professor Kay made possible carried with it his acknowledged credit of the important contributions that they made. This adds more than the reviewer can add to all those attributes which lie behind the personality of the senior author, who fortunately was able to see this work in final proof form before he was taken from his international family of scientific friends.

MORRIS M. LEIGHTON

Erratum note to be pasted on page 275, No. 4:

As submitted in manuscript form and as it later appeared in galley proof, the title of Dr. Cooper's paper read "Stones River Equivalents in the Appalachian Region." It is regrettable that the title did not appear in this, its correct form, in the final printing. The author naturally wishes to have the correct title listed in future bibliographies.

—THE EDITOR

VOLUME LIII NUMBER 5

THE JOURNAL OF GEOLOGY

September 1945

UPLAND TERRACES IN SOUTHERN NEW ENGLAND

GEORGE F. ADAMS
Columbia University, New York City

ABSTRACT

Recognition of the forms present has been the basic difficulty in the interpretation of New England geomorphic history. Joseph Barrell's recognition of broadly developed multiple terraces of marine origin is shown to be erroneous, since the existence of the terraces was concluded from incomplete evidence and questionable logic.

Multiple facets, as recognized by Douglas Johnson, can be observed directly on suitable profiles and to some extent in the field. These real forms indicate that New England has undergone complex peneplanation. Such a history is entirely compatible with the hypothesis of regional superposition applied to adjacent Appalachian areas.

INTRODUCTION

The New England Upland long has been a testing ground for many concepts of cyclic and noncyclic denudation. It still is being used for that purpose, for after forty-five years of research there is disagreement as to its geomorphic history.

The broad rolling surface of the crystalline rocks in western Connecticut was cited by W. M. Davis[1] as a good example of an uplifted, dissected peneplain of fluvial origin. This interpretation was challenged by R. S. Tarr,[2] who suggested the alternate hypothesis of noncyclic beveling. Davis replied, showing that noncyclic beveling could not adequately account for the subdued upland.[3] Since that time there has been general agreement that the upland is the result of cyclic processes, although few geomorphologists accept Davis' original interpretation of the surface as a single peneplain warped down more steeply near the Connecticut shoreline.

The earliest modification of the single-peneplain concept was proposed by Joseph Barrell.[4] He suggested that, although a peneplain had once been formed, it had largely been destroyed by later marine and fluvial erosion. According to Barrell, the present surface is not one peneplain but a series of seaward-sloping, dissected marine planes stepping down to the sea at intervals of from 100 to 400 feet. These dissected terraces were more fully described in a later paper.[5]

[1] "Physical Geography of Southern New England," *Nat. Geog. Soc.*, Mono. I, No. 9(1895), pp. 269–304.

[2] "The Peneplain," *Amer. Geol.*, Vol. XXIII (1898), pp. 351–70.

[3] "The Peneplain," *Amer. Geol.*, Vol. XXIII (1899), pp. 207–39.

[4] "Post-Jurassic History of the Northern Appalachians," *Bull. Geol. Soc. Amer.*, Vol. XXIV (1913), pp. 690–91.

[5] Barrell, "Piedmont Terraces in the Northern Appalachians," *Amer. Jour. Sci.*, Vol. XLIX (4th ser., 1920), pp. 227–58, 327–62, and 407–28.

Barrell's hypothesis of multiple marine planation virtually denies the preservation of a broad upland peneplain in New England. Other workers have modified Davis' hypothesis of a single warped peneplain in a different manner. According to H. S. Sharp[6] and Douglas Johnson,[7] the upland consists in the south of an older resurrected peneplain, but farther north it is a younger surface which had never been covered by later sediments. Johnson[8] has recognized that the upland possesses additional complexities which have not yet been fully studied. He doubts the marine origin of certain terrace-like forms on the following grounds: (1) The faint inclination of the supposed cliff zones are unlike any authentic marine forms. (2) The supposed sea cliffs are more faintly inclined than many hill slopes not regarded by Barrell as terrace remnants. (3) Slopes similar in form to the supposed sea cliffs face inland, away from the ocean. (4) The terraces are not recorded on Mount Desert Island, Maine, where the situation is favorable for their retention.[9]

G. W. Stose[10] has returned to the original interpretation of Davis for the New Jersey–Pennsylvania region and holds that a single warped peneplain is present. His interpretation is based chiefly on stratigraphic evidence gathered from the adjacent coastal plain, which would presumably apply to near-by New England.

[6] "Physical History of the Connecticut Shoreline," *Conn. State Nat. Hist. and Geol. Surv. Bull. 46* (1929), pp. 34–50.

[7] "A Theory of Appalachian Geomorphic Evolution," *Jour. Geol.*, Vol. XXXIX (1931), pp. 497–508; *Stream Sculpture on the Atlantic Slope* (New York: Columbia University Press, 1931).

[8] "Appalachian Studies I" (abstr.), *Bull. Geol. Soc. Amer.*, Vol. XL (1929), pp. 131–32.

[9] P. 7 of ftn. 7, *Stream Sculpture* . . . . (1931).

[10] "Age of the Schooley Peneplain," *Amer. Jour. Sci.*, Vol. CCXXXVIII (1940), pp. 461–76.

H. A. Meyerhoff[11] suggests that the supposed resurrected pre-Tertiary peneplain near the Connecticut shoreline is, instead, a Miocene marine erosion surface with some lower marine benches of later date.

Such are some of the conflicting views now held concerning the cyclic history of New England. Part of this conflict originates from different methods of observation, which lead different observers to believe in the existence of different forms. The rest of the conflict arises from the fact that various interpretations have been offered for the same forms. It is the purpose of this paper to re-examine the New England Upland, to evaluate the methods of observation employed by different investigators, and, if possible, to determine what significant forms are actually present. When this has been done, it will be pertinent to examine those hypotheses which seek to explain the forms recognized. By following this procedure some resolution of conflicting ideas may be achieved.

The interpretation of broad upland surfaces cut across complexly folded crystalline rocks, like those in southern New England, not only should be made on the basis of local evidence but should be compatible with regional histories of adjacent areas.

The studies of Douglas Johnson,[12] Karl Ver Steeg,[13] and others indicate fairly conclusively that in the adjacent portions of the Appalachians the erosion surfaces are of fluvial origin. The low descent of monadnock ridges toward present water gaps, described by Ver Steeg,

[11] "Tertiary Marine Planation in the Piedmont and Southern New England" (abstr.), *Bull. Geol. Soc. Amer.*, Vol. XLIX (1938), pp. 1954–55.

[12] Ftn. 7, *Stream Sculpture* . . . . (1931).

[13] "Wind Gaps and Water Gaps of the Northern Appalachians," *N.Y. Acad. Sci.*, Vol. XXXII (1930), pp. 87–220.

indicates that shallow water gaps existed at these sites on the Schooley peneplain and that a later marine invasion was never instrumental in affording a sedimentary cover thick enough to permit shifting of stream locations.

Johnson has shown that the existence of wind gaps in line with southeast-flowing streams indicates a regionally superposed drainage which has undergone disruption on a large scale without interruption by marine transgression since Schooley time.

If the Schooley peneplain can be traced into the New England Upland, then it would seem reasonable that surfaces in New England should be of fluvial origin and that the hypothesis of regional superposition, as outlined by Johnson, should find application.

The studies of Joseph Barrell indicate that any fluvial peneplains which may have been formed in New England have been largely destroyed and replaced by a series of broad, shallow terraces cut by marine agencies. This hypothesis of marine terracing is evidently incompatible with that of regional superposition as applied to the adjacent Appalachians, for it is difficult to see how the region immediately west of the Hudson River could have had a fluvial history while at the same time the area immediately east had a marine history.

The present study is not merely an effort to demonstrate the presence or absence of marine terraces but also an attempt to judge between two hypotheses. If marine terraces exist in southern New England, then serious doubt must be thrown on the hypothesis of regional superposition as applied to this and the adjacent Appalachians. If forms in New England can be assigned a simple fluvial origin, then further support is added to the hypothesis of regional superposition.

LOCATION OF THE AREA STUDIED

The present examination of surface forms has largely been confined to the Western Upland of Massachusetts and Connecticut (Fig. 1). The area is roughly bounded on the north by the Massachusetts-Vermont border, on the east by the Connecticut Triassic Lowland, on the south by Long Island Sound, and on

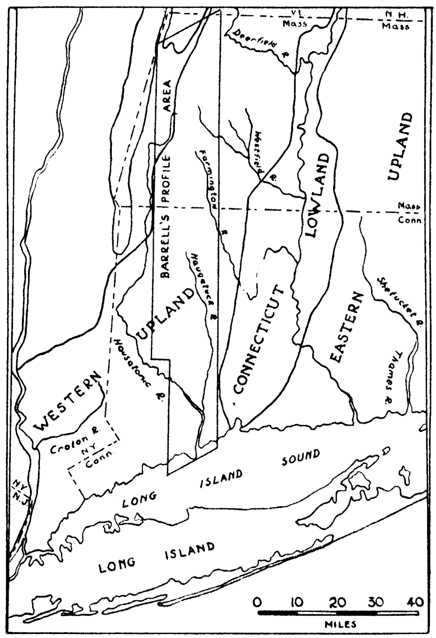


FIG. 1.—Index map

the west by the New York State line. In this area all the forms involved reach their most typical development. Here the best remnants of Barrell's supposed terraces are located. If they can be shown to indicate marine planation, then studies may be extended to the Eastern Upland, where, according to Barrell, terraces were weakly developed or poorly preserved. If the terraces do not stand analysis in the Western Upland, there is

little chance of finding critical evidence elsewhere in New England.

The surfaces described by Douglas Johnson are also well developed in the Western Upland. But since they extend southwest to the Manhattan prong of the New England Upland and eastward to the Eastern Upland, it will be necessary to consider those areas where additional data are required. The most intensive field work by the writer has been confined to the Western Upland.

THE MARINE-TERRACE HYPOTHESIS

The recognition of a series of terraces, cut in the crystalline rocks of the New England Upland and descending steplike to the sea, is the basis for Barrell's hypothesis of alternating marine and fluvial erosion. Although Barrell's field studies were far from complete at the time of his death in 1919 and many details of his postulated history were tentative, he and certain of his colleagues were convinced that the available data were sufficient to demonstrate the marine origin of the New England terraces. Accordingly, Barrell's abstracts, notes, and partially completed manuscripts were edited by H. H. Robinson and published in 1920.[14]

Other investigators[15] agree that New England and the adjacent area exhibit multiple terraces. Some variation is introduced in connection with higher terraces, which are generally considered to be fluvial in origin. None of these investigators attempts a critical analysis of the data upon which their conclusions rest. A paper by J. L. Rich[16] appears to accept Barrell's terraces but offers a somewhat different history to account for their preservation.

In view of the importance assigned to these supposed terraces, it appears necessary to subject the evidence for their existence to a more careful scrutiny than has heretofor been attempted.

DEVELOPMENT OF BARRELL'S HYPOTHESIS

Barrell, working on the physiography of the Housatonic folio in 1909–10, did not recognize marine terraces but interpreted the landscape in terms of the upland and lowland peneplains of Davis. In 1911 field work in western Connecticut and southwestern Massachusetts led to the same interpretation. Barrell was, however, on the watch for a Cretaceous shoreline, since he had presented evidence that Cretaceous beds had formerly extended farther inland than their present outcrop in adjacent areas. Seeking to determine the elevations of the accepted peneplains and possibly to find the expected shoreline, Barrell[17] constructed a projected profile covering a broad zone in southwestern Connecticut. This profile showed a pattern which could be interpreted as a series of steps or terraces. Extension of similar profiles

[14] Barrell, "Piedmont Terraces in the Northern Appalachians," *Amer. Jour. Sci.*, Vol. XLIX (1920), pp. 227–58, 327–62, and 407–28.

[15] Laura Hatch, "Marine Terraces in Southeastern New England," *Amer. Jour. Sci.*, Vol. XLIV (1917), pp. 319–30; Florence Bascom, "Cycles of Erosion in the Piedmont Provinces of Pennsylvania," *Jour. Geol.*, Vol. XXIX (1921), pp. 540–59; E. B. Knopf, "Correlation of Residual Erosion Surfaces of the Eastern Appalachian Highlands," *Bull. Geol. Soc. Amer.*, Vol. XXXV (1924), pp. 633–88; A. M. Pond, "Preliminary Report on the Peneplanes of the Taconic Mountains of Vermont," *16th Ann. Rept. Vt. State Geol.* (1927–28), pp. 292–314; H. A. Meyerhoff and M. Hubbell, "The Erosional Landforms of Central Vermont," *16th Ann. Rept. Vt. State Geol.* (1927–28), pp. 315–81.

[16] "Recognition and Significance of Multiple Erosion Surfaces," *Bull. Geol. Soc. Amer.*, Vol. XLIX (1938), pp. 1695–1721; see especially pp. 1714–17.

[17] "Geologic History of the Archean Highlands of New Jersey" (unpublished Master's thesis, Yale University, 1896), chap. viii, Part II.

over the whole western upland led Barrell to the conclusion that eleven benches, all of marine origin, could be distinguished.

The work thus started in New England was extended to the Maryland Piedmont and Coastal Plain. Here Barrell, using similar profiles, found similar benches which he correlated with the numerous unconformities in the coastal-plain sediments. The Piedmont benches trend parallel to the structure and could thus be explained by structural control. This was not believed to be true for the New England terraces, which were thought to be oblique to the structural trends, which are themselves oblique to the shoreline. Barrell found that the elevations of the New England terraces roughly coincided with those in Maryland, and from this he concluded that the structural control of the Maryland terraces was more apparent than real. He also believed that unconformities in the coastal-plain sediments could be used to date the terraces in both areas. On the basis of the terrace levels and unconformities, Barrell worked out the following tentative history for New England.

The sea advanced over the Cretaceous peneplain to the base of the Mount Greylock monadnock in Massachusetts, to cut the Becket terrace, now 2,400 feet above sea-level. This terrace was believed to be higher than the peneplain, which was considered to be partly preserved at an elevation of about 2,000 feet in an area east of the Hoosac Range and near the Deerfield River.

A slight emergence and subsequent stability resulted in cutting of the Canaan terrace, now at 2,000 feet.

A much greater emergence then caused the shoreline to retreat far to the south, somewhere beyond the present shoreline. The emergence was followed by mature fluvial dissection.

The sea then advanced over the maturely dissected land, which later remained stable long enough to permit marine agents to cut the Cornwall terrace, now at 1,720 feet.

The sea next retreated to the vicinity of the present shoreline, and streams again maturely dissected the land. Later submergence permitted marine agents to cut the Goshen terrace, now at 1,380 feet.

The terraces below the Goshen level are ascribed to successive emergences, with the possibility that the last four may have been due to eustatic changes. The names and levels are as follows: Litchfield, 1,140 feet; Prospect, 840 feet; New Canaan, 380 feet; Sunderland, 240 feet; Wicomico, 120 feet. The elevations given are for the inner margins of the terrace treads. All the terraces are described as sloping seaward between 2 and 8 feet per mile.[18]

It is important to remember that many of the terraces were supposedly formed after submergence of a maturely dissected land mass. Wave-cut terraces should, under these circumstances, have been developed first on islands and headlands. If the marine cycle progressed to maturity, the terraces should have become more continuous. But any such continuous terraces, if they ever were formed in New England, must since have been badly cut to pieces by stream dissection. In the nature of the case, therefore, it is impossible to determine whether the terraces were continuous at the time of their formation. Whether or not there was original continuity, the terraces today must be recognized, if at all, by remnants which are small in proportion to the intervening areas.

[18] Barrell, pp. 413–16 of ftn. 5 (1920).

If remnants of this kind are to be used to fix the positions of former shorelines, it must be demonstrated that they are of marine origin. A fair test of the marine origin of cliffs and benches may be applied by considering the effects of marine erosion on present shorelines and the results of later elevation of these features. With a clear idea of what should be expected to demonstrate the marine origin of cliffs and benches, it will then be appropriate to consider evidence for New England.

CRITERIA FOR THE RECOGNITION OF ELEVATED MARINE CLIFFS AND BENCHES

Elevated marine shorelines, developed on a formerly submerged sloping upland of complexly folded crystalline rocks, might be recognized by the following features:

1. *High-level benches which abut against steep cliffs.*—Such benches may be used if they bevel the structure and have considerable areal extent. Small local benches abutting against steep cliffs and beveling structure may arise from numerous fortuitous causes, such as jointing, exfoliation, slumping, et cetera—common in regions of complex structure. Minor benches, therefore, have no value as shoreline indicators. If large benches have been formed but later have been cut up by stream erosion, and the resulting hills are rounded by weathering, they are without value as indicators of former sea-levels, since there is nothing left to distinguish such hills from other rounded hills in the region. Isolated benches far from the supposed sea cliffs cannot fix the elevation of a former shoreline, since the bench may have been developed at an unknown depth below the supposed marine level, depending on the slope of the marine profile of equilibrium. All benches should slope seaward if they are of marine origin and have not been reversed in slope by subsequent warping.

It must also be shown that neither the bench nor the adjoining cliff has had its development controlled by structure or changes in rock type. Benches and cliffs attributable to such origin must exhibit specific marine features (undoubted wave-cut notches, sea caves, stacks, or marine deposits) before they can be used as shoreline indicators.

In judging the value of benches as evidence for former shorelines, the benches in question should be compared with others of similar size but so located as not to lend themselves to the marine interpretation. If it be found that the area contains similar or better benches at different elevations than those in question, or has similar benches facing away from the sea, or occurring in narrow estuaries, or sloping landward, then the value of the original benches as indicators of former shorelines is doubtful. This must be true when there is nothing to distinguish the benches attributed to marine action from others in the area which cannot be so interpreted.

If a broad, seaward-sloping bench is present, it should not be separated from the adjoining cliff by a sharply localized depressed area, unless such areas are underlain by weak rocks. Otherwise, it is highly improbable that the inner margin of a marine bench would be separated from its adjacent sea cliff by later fluvial processes before other parts of the bench were destroyed. The inner margin of a wave-cut bench normally has a steeper slope than parts farther out, is veneered with marine sand or gravel, and, when the sea withdraws, is likely to be protected by talus and slopewash from the sea cliff. Under these conditions streams should have great difficulty in

isolating the bench from the cliff unless the line of contact follows some structural or compositional weakness. If a weak zone is present, specific marine erosional features must be found before one may use the bench as an indicator of a former shoreline. Numerous depressions separating cliff from bench, where no marked structural weakness is present, are strong evidence against the marine hypothesis.

2. *Cliffs whose marine origin is adequately attested by the presence of wave-cut notches, sea caves, and stacks.*—Weathering and erosion may in time destroy notches and sea caves in a true marine cliff. In such event the marine origin of the cliff is difficult to demonstrate. This is especially true if the region exhibits abundant similar cliffs whose bases are at many levels different from those attributed to marine erosion or in positions not exposed to effective wave attack. Hills interpreted as stacks or offshore islands should rise from seaward-sloping benches.

CORRELATION OF TERRACE REMNANTS

If in a given region there are benches and cliffs which measure up to the standards enumerated above, it may fairly be concluded that marine erosion has modified the region to the extent indicated by the specific remnants. If the remnants are closely spaced areally and if their inner marginal elevations are accurately determinable, it may be possible to connect adjacent remnants to establish a former sea-level, if not a former shoreline. Such former sea-level positions may now be either horizontal or inclined, depending on whether crustal warping has affected them.

If the marine-terrace remnants are widely spaced, under no circumstances is it permissible to join such remnants with horizontal lines to support the interpretation that the terraces were once continuous and are oblique to the structure. Such lines are drawn on the basis of two assumptions: first, that the terraces were formerly continuous and, second, that they were uplifted without warping. Since neither of these two assumptions can be verified, any such lines must be considered as purely imaginary and as having no known relation to topographic or geologic elements. These lines, being imaginary, cannot be used as supporting evidence for the marine hypothesis, even if they do run oblique to structural trends. The marine hypothesis must be judged solely on the basis of the individual remnants of benches and cliffs.

If the remnants are widely spaced, it should be evident that many different correlations of the individual remnants may be possible under various assumptions of differential uplift. And if many correlations can be made, none of the restorations is a safe indicator of former sea-level, much less of the position of a former shoreline.

With these considerations clearly in mind, we may proceed to examine the evidence for the marine hypothesis in southern New England.

ANALYSIS OF LABORATORY EVIDENCE

Barrell's studies succeeded in emphasizing one point upon which there is general agreement: that the upland is not a smoothly sloping, dissected plane. That fact is amply attested by profiles and by field observations. Not only do the regional slopes vary in direction and amount, but there are many localized undulations in areas where the regional slope is fairly constant. Before any of these local undulations are accepted as remnants of marine terraces, it must be

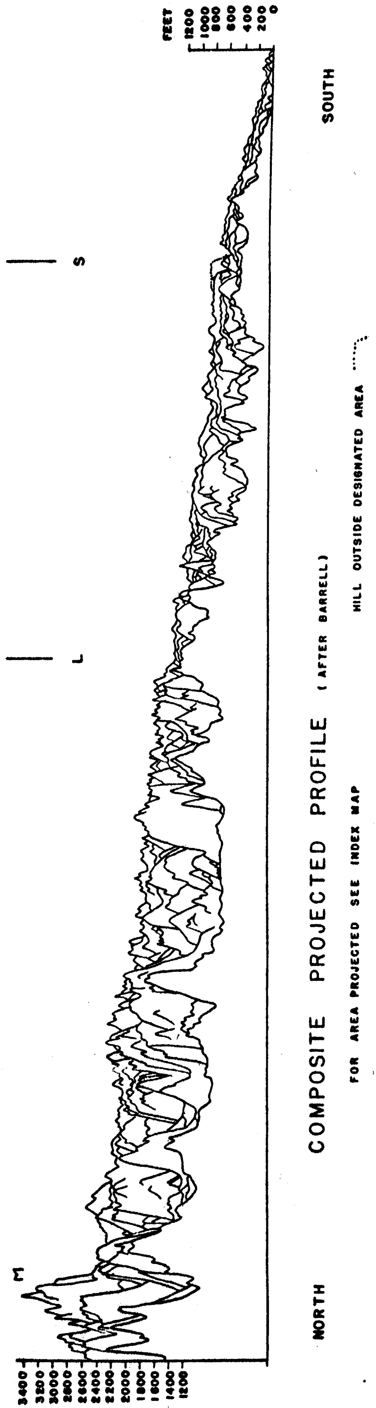


FIG. 2.—Composite projected profile (after Barrell)

shown that specific evidence exists which admits of no other interpretation.

Since the marine-terrace hypothesis as applied to New England was born from the study of maps and profiles, an analysis of this method of study is of primary importance.

The regional profile used by Barrell was of the projected type and covered an area in Massachusetts and Connecticut from one to one and one-third quadrangles wide, extending from Mount Greylock, Massachusetts, to Long Island Sound (Figs. 1 and 2). Such a single projected profile will be called by the writer a "composite projected profile." The method of its construction is thus described in the text of Barrell's paper and in an editorial note by Robinson:

A strip of cross-section paper is fastened parallel to the lower edge of a drawing board and the topographic sheet is fastened above so that the direction of view is at right angles to the length of the section and thus parallel to a T square placed against the lower edge of the board. The area to be projected is outlined and the contours which cut the front edge of the block are projected first to give the foreground. . . . . Then the topography beginning at the front and working toward the back of the block is systematically projected onto the cross-section paper with the aid of the T square. Each successive belt of topography *shuts out all behind it*[19] except the higher elevations, thus indicating what features are to be projected as the work progresses. . . . .

[EDITORIAL NOTE (BY H. H. ROBINSON): It should be said in regard to projected profiles that they are objective with respect to summit elevations but that the slopes may or may not be truly shown. There is room to *exercise considerable judgment* in the selection of slopes. The case may also arise where it is *advisable to omit a prominent object* in the foreground because it hides too much of the country behind, or if it seems *desirable to retain it*,[20] the topography hidden by it may be indicated by a dotted

[19] Italicized by the present writer.

[20] Italicized by the present writer.

line. A projected profile . . . . calls for the exercise in some degree of the pictorial sense].[21]

The method outlined above is characterized by H. H. Robinson, the editor of Barrell's posthumous paper, as being "more free than others from the subjective defect of picking and choosing the facts." This is far from true, for several reasons. First, a direction of view is chosen "which is best adapted to show the character of the culminating upland surface, as to whether it is a plane or a series of planes." The fact that the investigator chooses a particular direction of view indicates the subjective basis of the procedure. He must have an opinion as to the presence of terraces and their trend before he can select the direction of profiling that will best reveal their character.

The statement that a test is being made between a "plane or a series of planes" is misleading. Actually the question is whether the surface is a peneplain or a series of planes. Purely local irregularities of a tilted fluvial peneplain may well appear as terrace-like forms where the direction of view is normal to the slope of the surface (Fig. 3). Profiles made in this direction are inadequate to distinguish between a sloping peneplain and a series of planes. They must be supplemented by other profiles, looking up the slope of the surface, before one can distinguish between local irregular swells on a peneplain and extensive, systematic terrace remnants.

Another subjective element in the composite projected profile is the retention or omission of hills according to "desirability." If such a choice is permitted the investigator, it seems inevitable that he must tend to omit hills which do not fit into the terrace picture and which obscure hills farther away that would fit into the picture. Certainly, this is "picking and choosing the facts" to be represented as of significant value. A really objective profile would show the upland hills without selection and would be constructed without any single hypothesis in mind.

The subjective nature of the composite projected profile is further indicated when it is noted that Barrell found it necessary to go beyond the area outlined for profiling to take in certain hills the inclusion of which strengthens the evidence for one of the terrace levels (Fig. 2). The need for going outside the stated area (a whole quadrangle) to find support for a terrace level implies not only that the evidence for that terrace is weak but also that a single hill or group of hills may, on the composite pro-

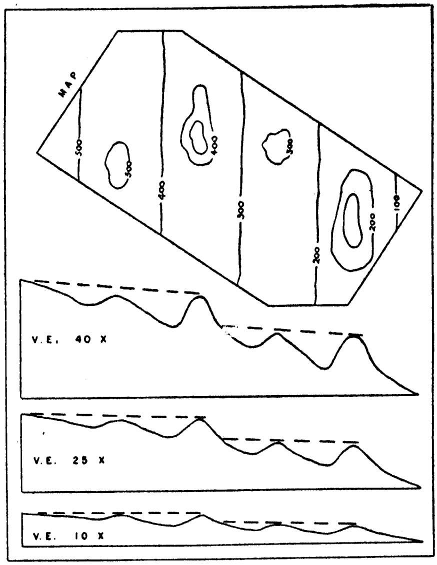


FIG. 3.—Composite projected profiles of scattered hills on a sloping plane, using vertical exaggerations of 40, 25, and 10 times (Barrell's method of projection). Dashed lines indicate the position of apparent dissected terraces. The hill slope appearing as the terrace riser seems to steepen as the vertical exaggeration is increased.

[21] Barrell, pp. 243–44 of ftn. 5 (1920).

file, give an appearance of terracing not supported by the topography as a whole.

An additional reason for doubting the efficacy of the composite projected profile is the dependence of the investigator on a single major projection. If errors in drafting are made, they may lead to erroneous interpretations upon which no check is available. In redrawing Barrell's profiles the writer found one or two errors, which, although not serious, were sufficient to demonstrate the dangers inherent in the composite projected profile.

A final deficiency in Barrell's type of profile involves the method of projection. Since the map is placed at an angle to the cross-section paper, distances in some cases are foreshortened. This results in a steepening of slope beyond that implied in the stated vertical exaggeration. Such steepening tends to favor the terrace hypothesis by increasing the apparent slopes of hills interpreted as terrace risers (Fig. 3).

For the reasons stated above, the writer is forced to conclude that the composite projected profile used by Barrell gives a distorted, incomplete, and dangerously subjective representation of the surface. By itself, such a profile indicates only certain topographic inflections, which may or may not be the product of terracing.

Barrell apparently realized some of the limitations of the type of profile he employed, for he supplemented the profile by tracing the supposed terrace levels across topographic maps of the region. This was done by connecting with continuous lines the inner margins of specific isolated "terrace remnants." The resulting regional terraces are shown in generalized form only, on a small index map (Fig. 1, Barrell). Unfortunately, terrace remnants are specifically named and located for but one small area. The other remnants must be searched for on topographic maps. By comparing these maps with Barrell's index map it has been possible to find most of the hills interpreted by him as terrace remnants, although some seem to be out of place and others could not be found.

It soon became apparent that the index map gives a false impression of the relative size and spacing of the supposed terrace remnants. This is because on the index map the symbols for the remnants are generalized and drawn on a greatly magnified scale, in order to make them clearly legible. Actually the remnants are very small in most cases, and the distances between them very great (Fig. 4, *A* and *B*).

As shown in the previous section, it is not permissible under any circumstances to assume that widely spaced terrace remnants can be joined by a line to indicate a former shoreline. Only when the marine origin of the remnants has been demonstrated and the proper correlation of these remnants has been fully established can a line connecting them be interpreted as marking a former margin of the sea. Barrell's strongest argument for the marine origin of his supposed regional terraces was that they cut obliquely across belts of varying resistance. But it should be fully appreciated that there are neither terraces nor terrace remnants traversing the lithologic belts of the New England Upland. The only things traversing those belts obliquely in a northeast-southwest direction are the lines of interpretation drawn by Barrell to connect isolated and widely spaced hill slopes regarded by him as terrace remnants.

EXAMINATION OF THE SUPPOSED MARINE-TERRACE REMNANTS

Since lines connecting supposed terrace remnants have significance as former shorelines only in case the remnants

can be shown to be of marine origin and in case the proper correlation of remnants can be fully established, it is pertinent first to examine the remnants, in order to determine their probable origin. scarps and ridgetops which have been interpreted by Barrell as remnants of marine terraces.

Although several high-level benches appear on Barrell's profile, only small

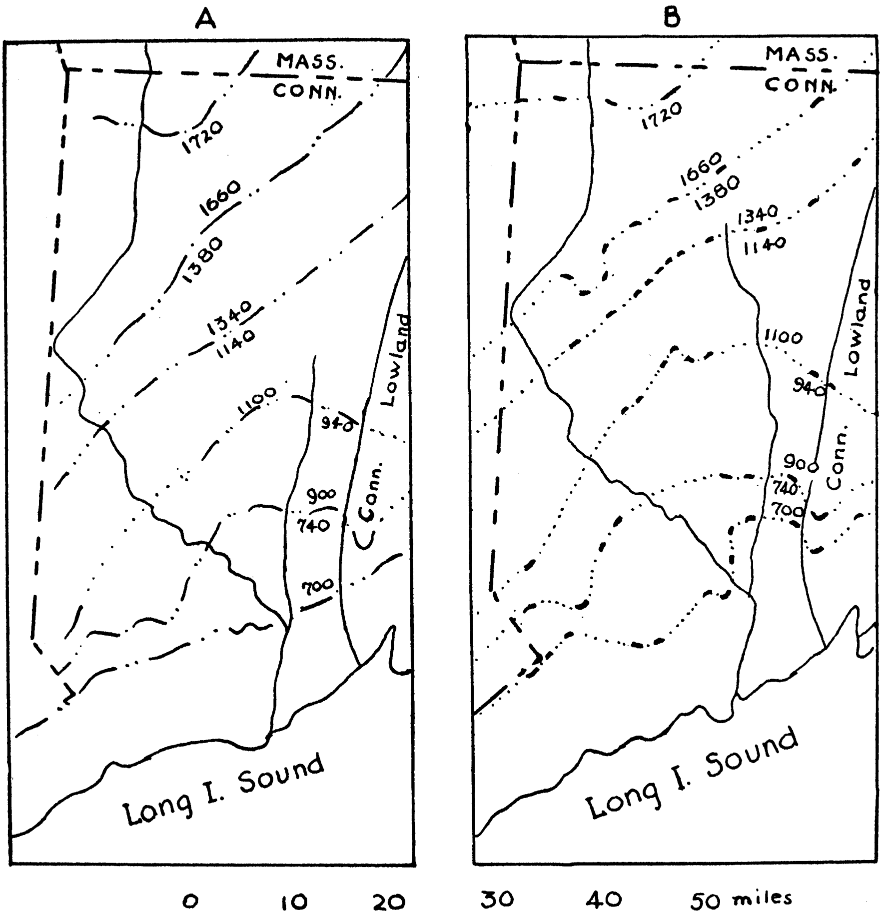


FIG. 4.—(*A*) Terrace restorations, redrawn from Barrell's "Piedmont Terraces," Fig. 1 (p. 247 of ftn. 5 [1920]). Long dashes represent stretches where the supposed terrace remnants are closely spaced but not necessarily continuous. Dots represent stretches where no remnants can be found. (*B*) The present writer's detailed plotting of supposed terrace remnants, taken from topographic maps.

The marine hypothesis can best be tested by applying the standards set up in a previous section to determine the marine origin of elevated cliffs and benches. We then turn to an examination of certain areas of two supposed terrace remnants are illustrated with a map.[22] This map, reproduced here as Figure 5, *A*, shows a portion of the Goshen terrace believed

[22] *Ibid.*, p. 255, Fig. 6.

by Barrell to have its inner margin now at an elevation of 1,380 feet and a portion of the Cornwall terrace with an outer margin now at 1,680 feet. Shaded to represent terrace remnants are all areas between 1,300 and 1,380 feet and between 1,600 and 1,680 feet. Hill slopes ranging through 80 feet of altitude are thus included. No known marine terraces have such surface relief. If it be assumed that the hills represent dissected remnants of a terrace, only the higher hilltops, if any, should be shaded as parts of the terrace. Or if the hills are interpreted as unconsumed remnants rising above the terrace level, only the lower land around them should be depicted as representing the terrace. When the supposed terrace remnants are thus restricted and are plotted at the assumed marine levels with due respect for the seaward slope postulated by Barrell, the areal extent of the supposed Goshen "bench" remnants is so enormously diminished as to cast grave doubt on their significance (Fig. 5, *B*). The same must be true in even greater measure for other "bench" remnants less clearly developed.

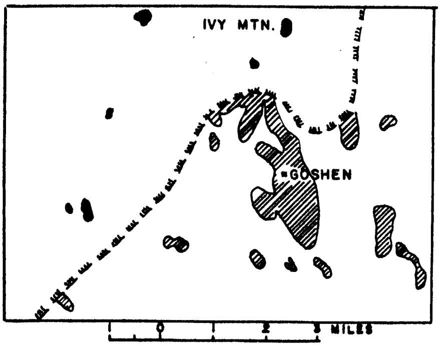


FIG. 5, *A*.—Map showing remnants of the supposed Cornwall terrace between 1,600 and 1,680 feet (*solid black*) and of the supposed Goshen terrace between 1,300 and 1,380 feet (*diagonal ruling*). After Barrell's Fig. 6 (p. 255 of ftn. 5 [1920]).

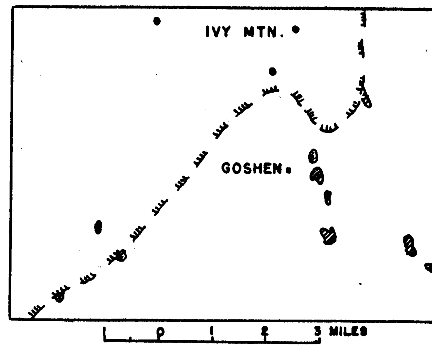


FIG. 5, *B*.—Map by the present writer showing actual summit areas at the elevations selected by Barrell as being those of terrace treads. Symbols same as in Fig. 5, *A*.

The remaining areas, interpreted as remnants of the Goshen terrace, nowhere abut against steep cliffs but are separated from the nearest scarps by valleys cut below the supposed marine level. Furthermore, the hills regarded as terrace remnants are not flat-topped. They are gently rounded and cannot be distinguished from other near-by hills not regarded as preserving marine levels. Two long, gently rounded ridges, one of which abuts against Ivy Mountain, Connecticut, suggest a level *100 feet higher* than the supposed Goshen level (Fig. 5, *A*). Examination indicates that this intermediate level, like the levels interpreted by Barrell as marine terraces, is only locally developed and cannot be traced across the upland to correlate with similar features. It thus appears that in the Goshen area, especially chosen by Barrell to demonstrate good marine terrace remnants, the existing hilltops and scarps fail to show any features which can safely be accepted as evidence of marine erosion.

Unfortunately, other bench areas are not figured in Barrell's paper. The writer has sought examples of flat areas of large

dimensions abutting against steep cliffs, sloping seaward and not controlled by structure and rock composition. No such features have been found. There are broadly rounded hilltops near the hypothetical shorelines; but these are off the assumed levels, are in the wrong position, or slope in the wrong direction to fit the marine-terrace hypothesis. It should be realized that most of the so-called "flat" areas referred to by Barrell[23] or interpreted from contour maps are, in reality, broadly rounded hilltops which need not imply the former existence of any bench or terrace tread. Most of these "flats" are separated from the nearest scarps by valleys—a condition not expectable under the marine hypothesis, for reasons earlier discussed.

Consideration of the areas cited above, as well as of other areas reported in this paper, and extended study of topographic maps lead to the conclusion that none of the broadly rounded hilltops show features confirming the marine-terrace hypothesis. Many of the best examples are above or below the supposed terrace levels. Others slope inland, and still others are in localities which would have been cut off from effective wave attack. In general, there is nothing which distinguishes the supposed sea cliffs and benches from any other cliffs and benches not regarded by Barrell as of marine origin.

In addition to the map of the Goshen area presented by Barrell and discussed above, several composite projected profiles of small areas are included in Barrell's paper on piedmont terraces. These profiles were especially chosen from a large number to show the best available benches and risers. Before discussing the individual profiles, it should be recalled that individual profiles of small isolated areas are not sufficient to demonstrate the regional nature of terraces. At best, they can only suggest possible marine levels. Such supposed marine levels must be adequately checked by profiles of intermediate areas to determine with what consistency they are maintained.

Examination of the published profiles (Barrell's Figs. 2, 3, 4, 7, and 8) raises doubt as to whether the topography indicates terracing of a marine nature. Barrell's Figures 2, 7, and 8 show areas on the Fall Zone where the regional slope is fairly steep—about 50 feet per mile (Fig. 7, *A*, of present report). The terraces are stated by Barrell to be weakly developed, most of them being considered as "re-entrant" and therefore not likely to show good breaks in slope on the profile. It may well be asked why poorly developed marine terraces in re-entrant areas should be selected for citation, since it is generally agreed that the earliest wave attack is concentrated on headlands and that major terracing is accomplished here.

Barrell's Figure 3 (Fig. 7, *B*, of present report) shows an area just west of the Connecticut Lowland near the inner margin of the Fall Zone. At first glance, the rise in slope from the 700-foot to the 900-foot level appears to support the marine-terrace hypothesis, because the rise is steep and joins broad, flattish areas to the north and south. Comparison of the profile with map and field data demonstrates defects of the composite projected profile and indicates that in this area there is little support for the supposed marine terrace. A hill erroneously labeled "Huntington Hill" (actually an unnamed ridge east of Straitsville) is located in an area that Barrell found necessary to include in his general profile in order to support the

[23] *Ibid.*, pp. 252 and 256.

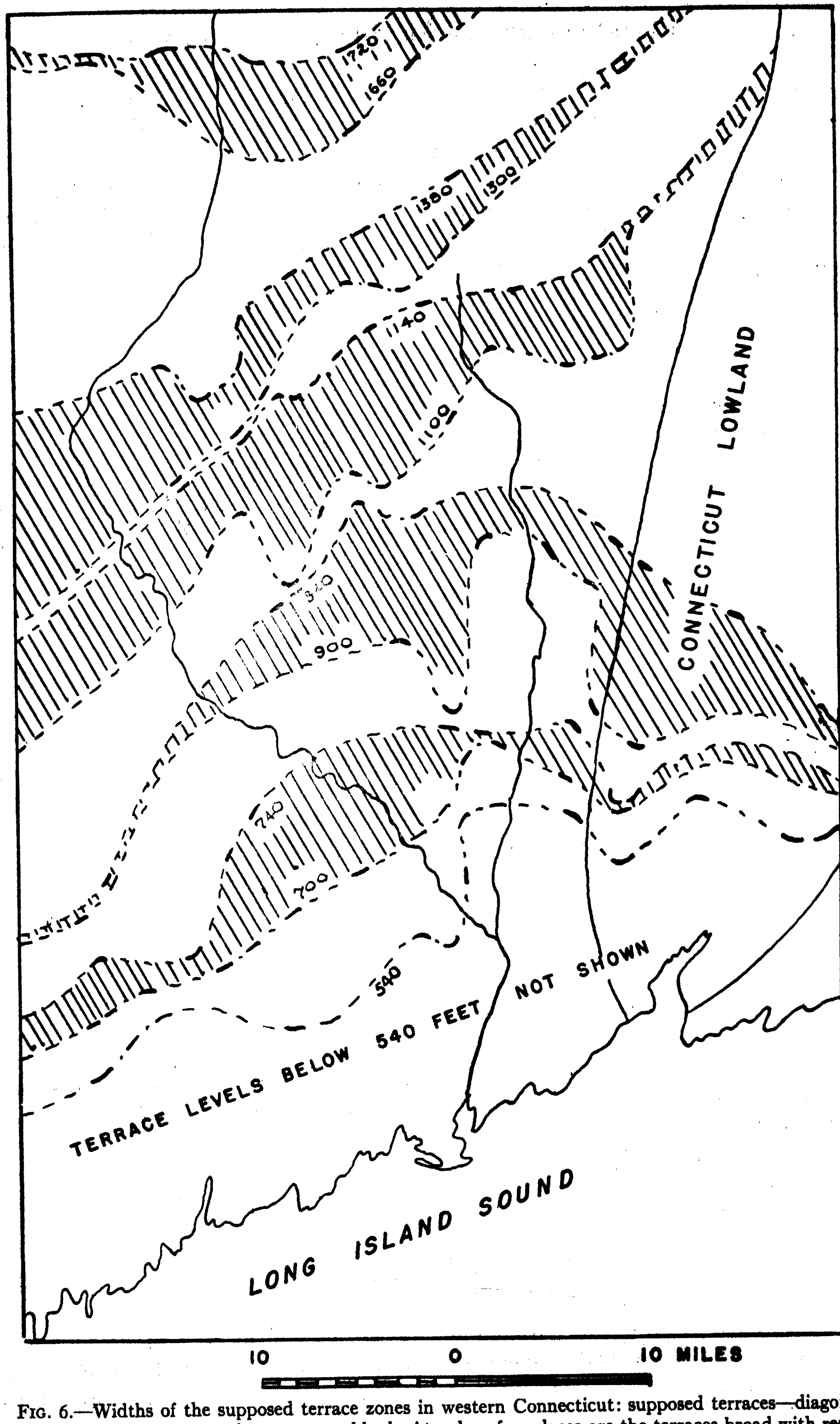


FIG. 6.—Widths of the supposed terrace zones in western Connecticut: supposed terraces—diagonally ruled; supposed risers separating terraces—blank. At only a few places are the terraces broad with respect to the separating risers. Riser slopes are predominantly gentle, as indicated by the great width of the riser zones.

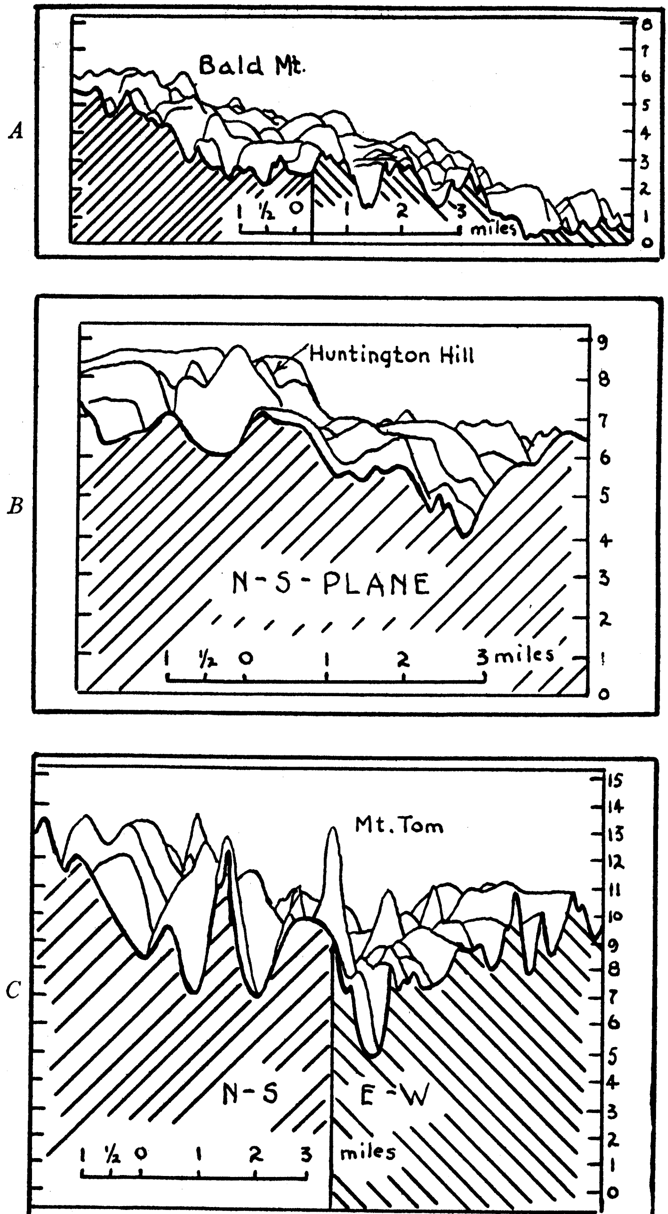


FIG. 7.—(*A*) Barrell's Fig. 2 (p. 252 of ftn. 5 [1920]). Projected profile of parts of the Stamford and Norwalk quadrangles, showing the Sunderland (200–240 feet), Pitcher Mount (280–320), New Canaan (340–80), and Appomattox (480–520) terraces. View looking N. 45° E.

(*B*) Barrell's Fig. 3 (p. 253 of ftn. 5 [1920]). Projected profile of parts of the Derby and New Haven quadrangles, showing the Towantic (700–740 feet) and Prospect (840–80 feet) terraces. View looking east. Huntington Hill is correctly located.

(*C*) Barrell's Fig. 4 (p. 253 of ftn. 5 [1920]). Projected profile of parts of the New Milford and Waterbury quadrangles, showing parts of the Litchfield (1100–1140 feet) and Goshen (1340–80 feet) terraces. View looking N. 55° E. (Captions taken directly from Barrell's figures.) Elevations given in hundreds of feet.

940-foot terrace level. As previously noted, the need for including this hill, lying outside the main area profiled, implies that the evidence elsewhere is weak. Were the hill to be omitted from the profile, the real Huntington Hill to the west, together with Woodruff Hill, would appear as gentle swells on the upland. Woodruff Hill is separated from the nearest 700-foot rounded ridge by a deep valley, clearly shown on the topographic map. The unnamed ridge east of Straitsville descends at its southern end to a small flat at 680 feet, which appears from the map to abut against a steep slope. Field examination shows that a deep valley, not shown on the map, separates the flat from the scarp. Huntington Hill drops gently down to a flattish ridge at 700 feet. In addition to the relations between steep and gentle slopes just discussed, it should also be noted that the lines on the profile indicating flat-topped ridges are misleading, since the ridges in question actually are not flat-topped but gently rounded.

It is significant that the higher hills in this region are located near the axes of plunging folds. Nothing distinguishes the seaward-facing slopes of these hills from others in the area. In fact, the general topography consists of strikingly curved ridge patterns, such as are normally developed by subaerial erosion of plunging folds. This aspect of the topography is in no way revealed by the composite projected profiles. In the absence of definite marine features, and for the reasons stated above, the apparent sharp break in slope shown on this profile cannot be considered evidence for a marine terrace.

Barrell's Figure 4 (Fig. 7, *C*, of present report) shows portions of the supposed Goshen terrace (1,380–1,340 feet) and the supposed Litchfield terrace (1,140–1,100 feet) in the vicinity of Mount Tom, Connecticut. Near the southern end of the profile there appears to be a reasonable amount of upland surface between 1,100 and 1,140 feet for a distance of 4 miles. Farther northward, elevations are lower; and, with the exception of Mount Tom itself, a broad hollow separates the 1,100-foot areas to the south of the hollow from those at 1,340 feet to the north. This hollow is a real feature, extending for several miles in a northeast-southwest direction, and is easily visible in the field. It is located where either a bench remnant at 1,140 feet or one at 1,340 feet should be present to support the marine-terrace hypothesis. Actually, the nearest rises on the north side of the hollow are several miles from the flattish areas at 1,100 feet to the south. Since there is so wide a separation of supposed sea cliff from supposed wave-cut bench, there is no possibility of demonstrating the marine origin of the forms present.

Mount Tom itself was considered by Barrell to be a stack or offshore island. Such features can be recognized only if they rise above wave-cut benches or have wave-cut notches at their bases. It has already been noted that Mount Tom rises from a hollow extending 100 feet or more below the supposed marine level. It cannot then be shown to rise from a platform of marine origin. Nor can wave-cut notches be observed in the field near the base of the mountain at the supposed marine level. All forms are of the normal subaerial type. Nowhere is there evidence of marine modification of subaerial forms.

If the considerations set forth on the preceding pages are valid, none of the profiles published by Barrell demonstrate the marine origin of forms in southern

New England. Since these profiles were selected by Barrell as the best among a larger number prepared by him, it must be concluded that the evidence offered in support of marine terracing in southern New England is far from convincing.

AN ALTERNATIVE METHOD OF VIEWING THE SURFACE

It may be urged that terraces would be visible if some better method of representing the upland surface were available. Such a method has been devised and applied to New England by Professor Douglas Johnson. This is the method of "multiple projected profiles," based on the construction of a large number of closely spaced projected profiles of belts of very limited width. Each profile shows the high points in a narrow zone, viewed from any direction the investigator finds convenient. After the profiles are drawn, they are transferred to stiff cardboard, cut out to show the relief, and then mounted on standards at intervals corresponding to the width of the zone. By this method a three-dimensional picture of the upland is made available.

Multiple projected profiles and the picture they present to the eye are almost entirely free from subjective influences. No particular orientation of the profiles is required, for, while a direction of view at right angles to the geologic structure or to the trend of terraces offers certain advantages, the final result, being in any case three dimensional, will reveal terraces if such exist, regardless of the direction selected. While individual profiles are being constructed, the projection of points is done in a purely mechanical manner and with no hypothesis in mind. There are few chances for error in construction or for omission of significant points. If occasional errors are committed, they are of minor importance, because the determination of forms present does not depend on a single line, as may be the case with a composite projected profile. Each profile of a multiple projected series is automatically checked by others on either side of it.

The great difficulty with multiple projected profiles is their cost in time, labor, and money. The writer has been spared this cost because, many years ago, Professor Johnson began experimenting with various kinds of profiles in order to discover the particular type, scales, and zoning best calculated to render an effective picture of the New England Upland. In the light of these studies and with the aid of grants from the Columbia University Fund for Research, Professor Johnson prepared two sets of profiles, each covering all of New England south of Vermont and New Hampshire. One set, trending north-south, was prepared by projecting parallel zones of terrain each 2 miles wide (Fig. 8). The other set, trending east-west, projects parallel zones each 1 mile wide (Fig. 9). These multiple projected profiles were placed at the writer's disposal.

Both sets of profiles have been studied by the writer, special attention being paid to the Western Upland, where the supposed marine terraces are considered by Barrell to be best preserved. This examination shows that, when the profiles are viewed in the same direction as that used in the construction of Barrell's composite projected profile, breaks in slope corresponding to those selected by Barrell can be seen. But when the multiple projected profiles are viewed from different directions and from different elevations, no continuous breaks in slope and nothing resembling remnants of terrace treads can be traced across

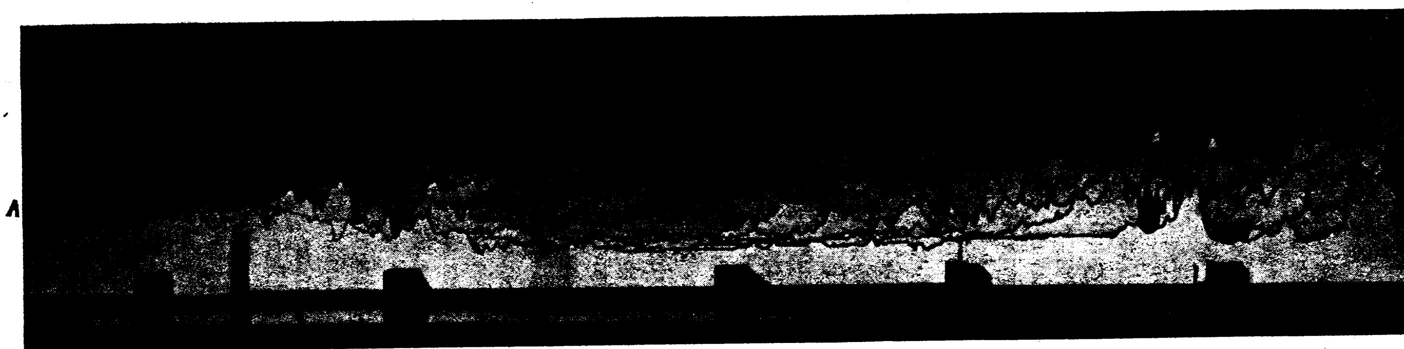


FIG. 8.—Multiple projected profiles (by Johnson), north-south set. Each profile covers a zone 2 miles wide. *ABCD*—Fall Zone; *EF*—inner margin of the Litchfield surface. Connecticut Lowland in foreground; Long Island Sound at left; Mount Greylock in right background.

the upland. The general impression gained from the multiple profiles is one not of terracing but of a subdued surface with irregular undulations. Many of the swells having linear extension trend oblique to Barrell's terrace trends, while many of the flatter areas do not slope seaward. The multiple projected profiles reveal many elements of form which do not appear on the composite projected type, but they fail to confirm the existence of regional terraces of marine origin. But it would have been premature to discard the marine-terrace hypothesis simply from the study of topographic maps and their derived profiles. This is especially true where the maps themselves are known to be somewhat inaccurate representations of the topography. It was recognized that marine features might exist which were lost in the mapping.

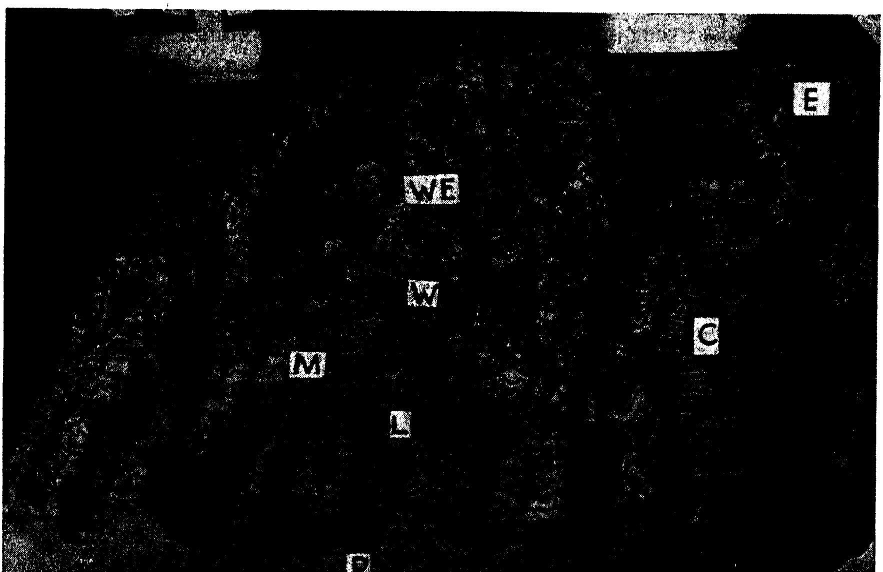


FIG. 9.—Multiple projected profiles (by Johnson), east-west set of western Connecticut and Massachusetts north of the Fall Zone, looking north. *C*, Connecticut Lowland; *W*, Western Upland; *E*, Eastern Upland; *P*, Pomperaug Valley, Triassic; *M*, Mount Tom, Connecticut, hollow; *W*, Winsted, Connecticut, hollow; and *L*, Litchfield surface.

istence of marine terraces in New England. This led to Johnson's early conclusion that the supposed marine terraces were nonexistent.

ANALYSIS OF FIELD EVIDENCE

After a critical study of topographic maps and composite projected and multiple projected profiles, the writer found that none of the evidence pointed to the

To guard against this possibility, the writer prosecuted field work in the Western Upland and in portions of the Eastern Upland in an effort to find evidence of marine terracing. Most of the work was done near the lines drawn by Barrell to represent margins of the supposed terraces, but other areas of possible marine cliffs and benches were studied where maps indicated their possible pres-

ence. Some sixty "remnants" along Barrell's lines were investigated.

None of these areas showed marine features such as sea caves, wave-cut notches, stacks, or marine deposits. None of the supposed sea cliffs could be differentiated from other scarps not regarded as sea cliffs by Barrell, whether on the basis of steepness, height, length of base, levelness of base, obliquity to rock structure, or contact with benches. Benches in these areas are small, and many slope in the wrong direction or are on the landward side of hills. Supposed bench areas are, in many cases, separated from the nearest scarps by marked depressions, some in areas where no depression is indicated on the maps. In brief, extended field examination adds nothing to the evidence in favor of the existence of marine terrace remnants in New England.

CONCLUSIONS ON THE MARINE-TERRACE HYPOTHESIS

On the basis of map, profile, and field study the writer concludes that the supposed terrace remnants postulated by Barrell on the basis of his composite projected profile are, for the most part, insignificant features of the landscape which do not extend far across the upland and that, even where linear topographic inflections extend for some distance, their marine origin cannot be demonstrated. The marine levels predicated by Barrell and the restored shorelines based on them do not appear to rest on an adequate foundation of recognizable facts. All the forms observed find full and normal explanation on the basis of a continuous fluvial history. The geomorphic history postulated by Barrell, involving alternate marine and fluvial cycles, must accordingly be regarded as of doubtful validity.

FACETS OF THE NEW ENGLAND UPLAND

While valid evidence of a group of marine terraces on the New England Upland is lacking, one may distinguish a number of clearly developed "facets" of the upland surface, sloping at different angles. Some of the facets are separated from neighboring ones by steeper slopes, which may or may not be significant of the origin of the facets. The existence of such facets was early recognized by Douglas Johnson[24] from his study of multiple projected profiles. Examination of Barrell's composite projected profile (Fig. 2) indicates that to the north of the point *M* a mountain group rises abruptly far above the adjacent upland level, perhaps as a monadnock mass surmounting the surrounding peneplain. Between *M* and *L* the culminating hills of the dissected upland are strongly undulating, with no definite indication of terrace forms. If terraces ever did exist in this area, they have been so cut to pieces by stream erosion that they cannot now be distinguished from irregular swells on the sloping peneplain. At *L* there is a decided steepening of slope, south of which the culminating hills extend, with only minor breaks in the more gentle regional slope, to the point *S*. While a few broad valleys in this sloping facet leave higher hills to the north rising above lower hills to the south, thus giving a crude suggestion of terracing at one or more points, no undoubted terraces exist. South of *S*, the regional slope is notably steeper; but indications of terraces, as admitted by Barrell, are even weaker than elsewhere. This more steeply sloping facet has long been recognized as the Fall Zone peneplain, which intersects the less steeply sloping upland peneplain (Schooley?) at *S*. The only clear indication of a regional terrace

[24] Pp. 131–32 of ftn. 8 (1929).

shown on Barrell's composite projected profile (Fig. 2) is the broad upland facet extending from *L* to *S*. It is a reasonably even surface of moderate slope, terminated landward and seaward by notably steeper slopes. It does not correspond to any of Barrell's supposed terraces, being a broader feature of great areal extent. The question of its origin will not be considered at this time.

The regional extent of Barrell's supposed marine terraces (Fig. 2) may be determined by extending the inner and outer margins of the supposed terraces across the upland. This may be done directly on topographic maps by plotting the hills which rise to the levels selected for the inner and outer terrace margins. The lines thus drawn should consistently be close together where they delimit the supposed terrace risers and notably farther apart where they delimit the supposed treads.

Using the terrace levels selected by Barrell, the writer has drawn lines representing the landward and seaward terrace boundaries for the Connecticut portion of the Western Upland, where the supposed terraces are most clearly indicated on the composite projected profile (Fig. 2). As shown in Figure 6, the lines are not consistently close together at the supposed risers, nor do they remain consistently far apart on the supposed treads. Instead, the lines wander irregularly over the upland. Such an analysis indicates that terraces, if present at all, are only locally developed and are separated by large areas where no indication of terracing is present. It thus appears that multiple terraces of regional extent do not exist at the elevations selected by Barrell, on the basis of his composite projected profile.

To guard against the possibility that regional terraces might exist and not show on a composite projected profile, the writer has used another method of exploring the upland for such terraces. This is to construct generalized contours at regular vertical intervals, so drawn as to eliminate the disturbing effects of stream dissection, and to examine these contour lines for indications of regional terracing. The generalized contours are drawn, starting at the first interval above sea-level, by joining those points on interstream areas which first reach the level being contoured. Although personal judgment must be used in determining the precise positions of the contours, the general positions are fixed by a large number of points for each line and hence may be accepted with some degree of confidence as showing the general form of the upland before dissection. The map thus prepared (Fig. 10) indicates that the method of construction is sound, since it shows the same major features as are observed on the multiple projected profiles. The map shows in a single illustration, easily reproduced, features which require for their appreciation several photographs of the profiles.

If regional terraces are present in the Western Upland, the generalized contours should show consistent bunching on the risers and wide spacing on the treads of any terraces present, provided that the contour interval is smaller than the differences in terrace elevations. Examination of the map indicates that consistent alternate bunching and wide spacing of the lines is not regionally developed. Local bunching may be observed, as, for example, southwest and northeast of Peekskill, where the upland surface descends more or less abruptly to the Triassic Basin and its northeastward extension, along the Ramapo fault-line scarp. Another scarp, trending north-south, west of the lower Housaton-

ic River and east of the New York boundary, appears to have been the southwestern side of a broad sag associated with the Housatonic River, the sag having been locally steepened by excavation of less resistant limestones. A third zone of bunching extends northeastward from a point some miles northeast of Bulls Bridge on the Housatonic River to the Farmington River. Other scarps are indicated in the northwestern portions of the upper Westfield and Deerfield River basins, where the scarps trend parallel to structure and may be explained by the excavation of weak rock on the basin side of the scarp. The only one of these

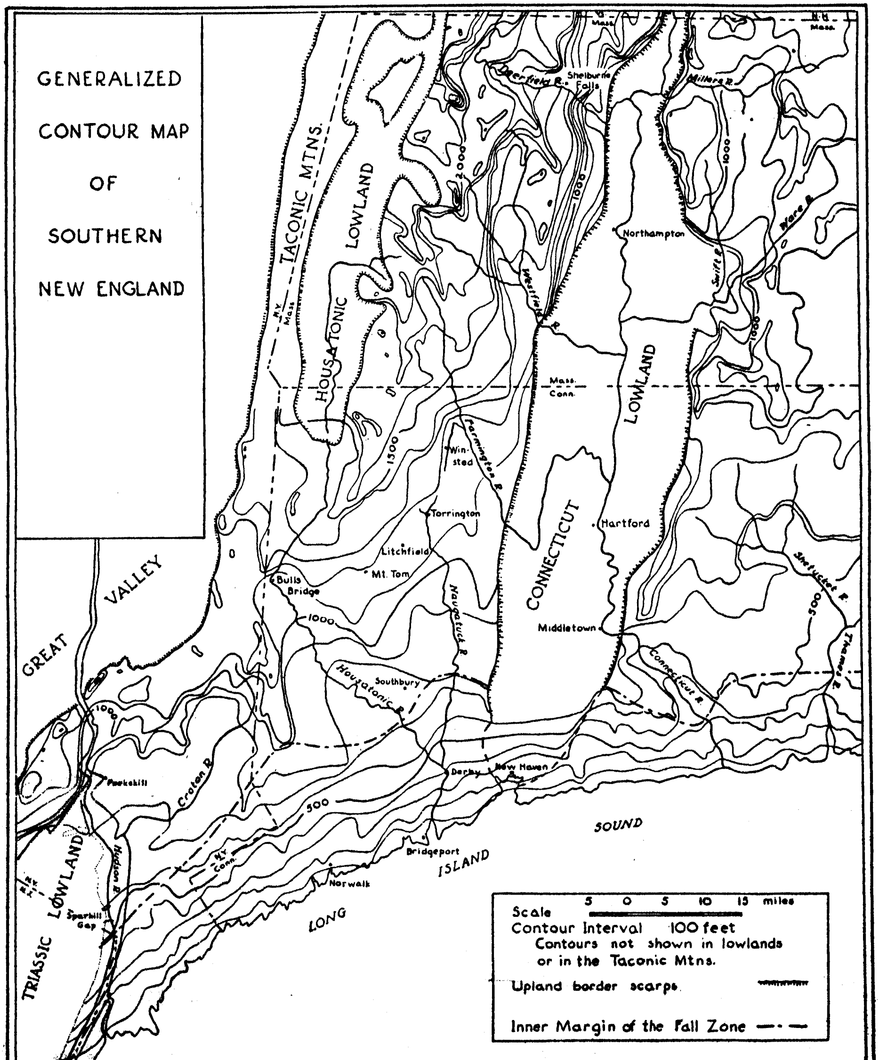


FIG. 10.—Generalized contour map of southern New England

zones of more closely spaced contours which seems unrelated to rock difference or known geologic structures is the one extending northeastward from Bulls Bridge to the Farmington Valley northeast of Winsted.

The relation of this local belt of steepening to the major facets of the Western Upland should be appreciated. Beginning at Long Island Sound, along the southern border of the Western Upland, a relatively steep slope, of 50 feet per mile, toward the south-southeast is maintained to about the 800-foot contour. For the next 20–30 miles northward the regional slope is almost due southeast and diminishes to 30 feet per mile. The facet of more gentle slope is gradually terminated northwestward by a narrower zone of relatively steep slope (60 feet per mile), shown on the map as the zone of bunched contours extending from the Housatonic River above Bulls Bridge and described above. To the northwest of this steeper zone, regional slopes generally diminish; but any exact determination of slope is rendered difficult by numerous local structurally controlled swells and hollows.

The writer concludes that the generalized contour map indicates a single surface or facet which with some show of reason might be called a "regional terrace." This is the surface sloping southeast at 30 feet per mile and terminated landward and seaward by more steeply sloping surfaces. It is shown on Barrell's composite projected profile extending from *L* to *S* (Fig. 2). This poorly developed terrace will be called the "Litchfield surface," from its best development in Litchfield County, Connecticut. The origin of the steeper zone, or "riser," terminating the Litchfield surface at the northwest is not clear; but Johnson[25] has suggested that such inflections of the present surface may represent monoclinal warping of a buried peneplain from which the sedimentary cover was later stripped. This interesting suggestion might be applicable if there were any indication of such a sedimentary cover over the upland surface. Johnson has presented other evidence that a sedimentary cover did extend over this area but that it blanketed the Fall Zone peneplain, which has since been beveled in this area to form the present upland surface. It appears reasonable that such beveling would have removed any structural inflections developed on the Fall Zone peneplain.

Examination of Johnson's multiple projected profiles indicates that multiple regional terraces are not present on the New England Upland. A few broad changes in slope are present, corresponding to those suggested by the generalized contour map and described above. The same structurally controlled irregularities are discernible.

Thus, after utilizing various methods of studying the Western Upland, it appears necessary to conclude that multiple terraces of a regional nature do not exist. Similar study of the Eastern Upland leads to the same conclusion. The single regional terrace which does appear as a real feature is confined to the Western Upland and is but faintly differentiated from the rest of the surface. It also appears from these studies that the upland of southern New England consists of a few broad facets, as earlier noted by Johnson, the significance of which will be considered in a later paper.

CONCLUSIONS

The more prominent upland irregularities used by Barrell to support the marine-terrace hypothesis are closely re-

[25] Pp. 7–8 of ftn. 7, *Stream Sculpture* .... (1931).

lated to geological structure and rock composition. Similar irregularities not appearing at the supposed terrace levels are numerous and are related to the same causes. Such irregularities are fully explained as the product of weathering and fluvial erosion and are to be expected in a peneplaned area of complex structure and diverse rock types. They doubtless existed in subdued form before uplift and dissection of the peneplaned surface. They have been emphasized where dissection is deep and where differential lowering of closely adjacent surfaces has occurred.

In view of this reasonable and adequate explanation of the upland irregularities, they cannot be accepted as remnants of marine terraces unless marine modification of subaerial forms can be established. No convincing evidence of such modification has been offered.

The writer agrees with Bryan[26] and his co-workers that it is impossible to prove that the sea did not make repeated advances over this area, to cut marine terraces. But, if such terraces were ever formed, they were so insignificant that weathering and fluvial erosion have since effectively obscured them or later fluvial peneplanation has completely removed them. No valid evidence that such forms ever existed has been found.

Applying this conclusion to the larger picture of Appalachian geomorphology, it can be said that no barrier exists to the application of the hypothesis of regional superposition in the New England area. Further search is, however, indicated to discover more positive evidence in favor of this hypothesis. This, too, will be the subject of a later paper.

[26] Kirk Bryan; A. B. Cleaves; and H. T. U. Smith, "The Present Status of the Appalachian Problem," *Zeitschr. f. Geomorph.*, Vol. VII (1933), pp. 312–20.

Acknowledgments.—The study presented in the preceding pages was suggested by Professor Douglas Johnson and has profited from his continuous critical supervision. The writer is further indebted to the searching criticisms of the Seminar in Geomorphology at Columbia University and to the friendly encouragement of Professor Bertram T. Butler, of the City College of the College of the City of New York.

SMALL-SCALE STRUCTURES IN THE COCONINO SANDSTONE OF NORTHERN ARIZONA

EDWIN D. McKEE
Museum of Northern Arizona, Flagstaff

ABSTRACT

Small-scale structures, including ripple marks, slump marks of two types, and rain patches, occur locally on the surfaces of steeply dipping laminae in the eolian Coconino sandstone. These surface features are described and analyzed. Comparisons are made between them and similar structures observed in the modern dunes of the Navajo country. Most of these structures were reproduced experimentally in a wind tunnel and sand box, where controlled conditions made possible a determination of factors involved.

INTRODUCTION

The Permian Coconino sandstone of northern Arizona, which is a highly cross-laminated deposit of clean, well-sorted quartz sand, generally considered to be of eolian origin, locally contains an abundance of ripple marks and other minor structures and, in some places, great numbers of fossil footprints. A desire to determine accurately the environment in which these features were formed has prompted an examination of comparable structures among modern dunes in the Navajo Indian Reservation and the conducting of a series of experiments, the results of which are presented in this paper.

Shortly after the conclusion of the first season of investigation the definitive work of R. A. Bagnold[1] on the physics of blown sand and desert dunes became available to the writer and was of inestimable value in completing the studies. The painstaking experiments conducted by Bagnold made unnecessary a further search for many facts required in obtaining a satisfactory explanation of various features involved. Certain of the experiments that had a bearing on the study of small-scale structures in the Coconino have been repeated at the laboratory of the Museum; additional ones have been made where more information seemed desirable.

TYPE OF DUNES

Evidence of the eolian origin of the Coconino sandstone has been presented in papers by the writer[2] and by Parry Reiche[3] and will not be repeated here. Recognition of the type of dune, however, is essential to a full understanding of the ripple marks and other minor structures forming the subject of this paper and therefore warrants discussion.

Among various dune classifications that have been proposed, those by J. T. Hack[4] and by Bagnold,[5] based on form, in so far as it reflects genesis, are considered here because of their simplicity and applicability. Hack recognizes three primary dune forms: (1) transverse (including barchan), (2) parabolic, and (3) longitudinal. He also refers to several special types, developed as a result of such obstacles as topographic irregularities.

[1] *The Physics of Blown Sand and Desert Dunes* (New York: William Morrow & Co., 1943).

[2] "The Coconino Sandstone—Its History and Origin," *Carnegie Inst. Wash. Pub. 440* (1933), pp. 112–14.

[3] "An Analysis of Cross-lamination: The Coconino Sandstone," *Jour. Geol.*, Vol. XLVI (1938), pp. 916–18.

[4] "Dunes of the Western Navajo Country," *Geog. Rev.*, Vol. XXXI, No. 2 (1941), pp. 240–41.

[5] Pp. 188–89 of ftn. 1 (1943).

Bagnold considers as true dunes only those that can exist independently of any fixed surface feature and lists two fundamental types: (1) barchan, crescentic, or transverse; and (2) longitudinal or seif. Other types of sand accumulation, caused directly by sand obstructions such as bushes, rocks, or cliffs, he refers to as "sand shadows" and "sand drifts." These include the parabolic dune of Hack, which results from anchorage by vegetation.

In the Coconino sandstone there are no indications, such as projections from an underlying surface, that the deposits, even at the base, represent sand accumulations controlled by topographic irregularities. Nor is there any evidence in the form of plant molds or bedding disturbed by roots to suggest that vegetation played a part, with resulting parabolic forms. Of the other principal dune types, it seems unlikely that the longitudinal dune is represented, for, as pointed out by Reiche,[6] such dunes develop in districts of slight accumulation, where the sand is chiefly in transport. Furthermore, structures typical of this type of dune, as described by Bagnold, have not been recognized in the Coconino sandstone.

What appear to be small crescentic or barchan dunes may be observed in various parts of the Coconino sandstone where exposures are suitable. The curving fronts illustrated in lamination planes and the extended wings with high-angle dips in divergent directions are very prominent locally. In this formation, however, most of the planes of erosion and sets of steeply dipping laminae appear to be on a far larger scale and, as pointed out by the writer,[7] probably were formed as transverse (straight-fronted) dunes—a type differing from the barchan largely in degree. In typical transverse dunes the scale is so great that plunging "anticlines" or "synclines" with arclike pattern in cross section are not easily recognizable. Instead, the long, gently dipping planes of erosion, and the thick series of steeply dipping laminae with constant direction are the conspicuous features. Most of the Coconino structures appear to be of this type rather than of the barchan variety.

SELECTIVE PRESERVATION OF LEE-SLOPE SEDIMENT

Sand-dune deposits preserved in the geologic column probably consist almost wholly of lee-slope material. This conclusion was reached by F. W. Shotton[8] on the basis of theoretical considerations, and the principle involved has been demonstrated and explained by Bagnold[9] through studies of modern dunes in the Libyan Desert. In general, deposits of the approach side are only temporary and sooner or later are moved over the crest, building new laminae to leeward. Exceptions, consisting of the relics of windward-slope deposits, are locally preserved, along with the dominant lee-side laminae, as pointed out by W. O. Thompson[10] and other investigators; but these windward-side deposits are usually relatively easy to recognize as such. In barchan and transverse dunes laminae of the lee-side dip at a high angle, approaching 32° or 33°, and in directions that are, in general, comparable to the dominant wind direction of the area. Windward-

[6] P. 911 of ftn. 3 (1938).

[7] "Structures in Modern Sediments Aid in Interpreting Ancient Rocks," *Carnegie Inst. Wash. Pub. 501* (1938), pp. 688–90.

[8] "The Lower Bunter Sandstone of North Worcestershire and East Shropshire," *Geol. Mag.*, Vol. LXXIV (1937), p. 544.

[9] Pp. 241–42 of ftn. 1 (1943).

[10] "Original Structures of Beaches, Bars and Dunes," *Bull. Geol. Soc. Amer.*, Vol. XLVIII (1937), pp. 747–50.

side deposits, on the other hand, dip in the opposite direction and at angles less than 12°.

Applying the preceding considerations to the problem of the Coconino sandstone, it is clear (1) that the lee-side deposits should in most cases be distinguishable from those formed to the windward through a consideration of the direction and the degree of dip and (2) that laminae formed on the windward-side should be relatively scarce in this formation. Reiche[11] in an analysis of the Coconino cross-lamination has demonstrated that this is the case. Out of many hundreds of readings of lamination dip and strike, a large majority were found to dip between 15° and 30° in a southerly direction. The few exceptions dipping opposite were low-angle surfaces and therefore have been interpreted as representing true windward-side deposits.

EVEN-SURFACED LAMINAE

In wind-deposited sand being accumulated through saltation there is a pronounced tendency, under most conditions, to develop ripple marks. Because of the bouncing process involved, initial surface irregularities are accentuated and mounds of sand pile up rhythmically beyond them. The results of this tendency are conspicuous on the windward surfaces of most dunes. Where flat, rather than rippled, surfaces have formed among such deposits, the cause, according to Bagnold,[12] may be (1) continuous, rapid deposition that precludes opportunity for sand-sorting; (2) increase in wind strength, causing the length of the ripple marks to increase to such an extent that they finally disappear; or (3) sand of irregular grading in which some grains are much larger than others and are too large to be moved by the available wind, thus tending to break up any orderly ripple advance.

In the Coconino sandstone most lamination surfaces are extremely smooth and even. An adequate explanation of this feature seems essential to an understanding of the formation. A few of the laminae dip with low angle in a direction opposite the normal and so are considered to be windward-side deposits. In these the lack of rippling is probably the result of very rapid accumulation, for, as explained by Shotton,[13] deposits of the windward side have a good chance of permanent preservation only where the wind is being furnished an excessive supply of sand. Such rapid deposition has been shown by him to be unfavorable to ripple development.

Most laminae in the Coconino appear to have formed on the lee sides of dunes and therefore must have rippleless surfaces for a reason fundamentally different from that applied to windward-side deposits. Instead of being developed by the saltation of sand grains, these laminae are brought about largely by mass sliding or avalanching of sand down the steep lee slopes, as pointed out by Bagnold;[14] hence there is normally no tendency toward ripple development.

The principle involved in lee-side deposition is demonstrated by an experiment suggested by Bagnold in which sieved sand, dyed a different color for each of three size grades, is used. A sample in which these three grades are well mixed is allowed to avalanche down the steep slope of a sand box. When the material settles at the angle of repose, it is clear from the appearance of color bands that each sand grade has separated out into a separate flat layer or lamina.

[11] Pp. 905–82 of ftn. 3 (1938).

[12] P. 153 of ftn. 1 (1943).

[13] P. 544 of ftn. 8 (1937).

[14] P. 239 of ftn. 1 (1943).

Only on the sides or wings of barchans where wind currents often cross the steep slopes is there more than an occasional opportunity for re-working lee-side deposits and developing ripple marks (Fig. 1). Thus, the general lack of ripple marks and of surface structures in the Coconino sandstone appears to be largely a function of selective preservation of certain portions of the dunes, in which, with minor exceptions, conditions of deposition are unfavorable for ripple development.

FIG. 1.—Wing of barchan dune, showing ripple surface and slump marks on lee side, Tuba, Arizona

RIPPLE MARKS

GENERAL FEATURES

Ripple marks are not abundant in the Coconino sandstone but locally are moderately numerous. Approximately fifty sets from localities near Seligman, north of Ashfork, and in Oak Creek Canyon, Arizona, were examined in connection with this study. Many of the specimens, unfortunately, were not *in situ*, having been discovered in slabs of sandstone quarried for use as paving stone; but enough sets were located and examined in the field to determine their original positions with respect to the cross-laminated layers. On all specimens studied, a record was made of the height of the ripple crests, distance between crests, the ripple index (defined as the wave length divided by the amplitude), the degree of parallelism of individual ripples, and the amount of rounding of the crests.

Coconino ripple marks are all of one general type. Analyzing the results of examination, five features are found to be characteristic. First, they have a large ripple index, the lowest of those measured being seventeen, and most of them ranging well above this figure. Sec-

ond, all specimens found *in situ* have crests trending up and down the surfaces of sloping laminae and in no case normal to the dip. Third, the ripple crests are definitely rounded, as illustrated in every specimen examined. Fourth, there is a concentration of coarser sand grains on or about the crests in many specimens. Fifth, the crests and troughs within each set are remarkably straight and parallel, showing little change of direction even on large surfaces where exposures are continuous for 3 or more feet.

RIPPLE LENGTH

The length (distance between crests) of most ripple marks in the Coconino sandstone is between 100 and 150 mm., although exceptions are not rare. The largest ripple mark to be examined, a specimen from near Ash Fork, Arizona, is 302 mm. in length. Two other specimens measured only 75 and 85 mm., respectively.

According to E. M. Kindle,[15] "the size or amplitude of wind ripple mark varies but little with variation of wind velocity . . . . the principal factor controlling the slight amount of variation which is shown by the amplitude, appears to be the coarseness of the sand." This conclusion, based on general observations, apparently is incorrect. Bagnold[16] describes in detail the lengthening of ripple marks as a result of increased wind velocity, independent of changes in grain size and sorting. Furthermore, this relationship has been demonstrated repeatedly during the present investigation under controlled conditions of a wind tunnel where grain size could be kept constant and wind velocity changed as desired.

The variations in wave length recorded for ripple marks in the Coconino sandstone probably represent differences in wind velocity. Grain size and degree of sorting are too similar in the samples studied to account for any large differences in ripple length such as are represented by the extremes noted above.

RIPPLE INDEX

High ripple index (above 15) is considered by Kindle[17] to be "the most striking characteristic of wind-made ripple mark," and studies by numerous other geologists have established the validity of using this feature in distinguishing between wind- and water-formed ripple marks. Bagnold[18] has shown that the index of wind ripple marks is controlled by the degree of sorting of the sand, very uniform grains tending to give only very low ripples and hence high indices. In using a single grade of sand (0.19–0.27 mm.) he found that "the ratio of height to wave length was usually 1 to 70, and never more than 1 to 30." Similar experiments conducted by the writer gave the results shown in Table 1.

TABLE 1

	Average Index
Mixed sand (0.5–1.0, 0.25–0.5, and 0.125–0.25 mm.)	16.0
Fine-grained sand only (0.125–0.25 mm.)	27.0
Medium-grained sand only (0.25–0.5 mm.)	26.0
Coarse-grained sand only (0.5–1.0 mm.)	29.0

Ripple-index determinations have been made for twenty-one specimens in the Coconino sandstone. Unfortunately, accurate measurements could not be made in other available specimens because ripple crests were weathered. The

[15] "Recent and Fossil Ripple-Mark," *Can. Dept. Mines, Geol. Surv. Mus. Bull. 25* (1917), p. 12.

[16] P. 151 of ftn. 1 (1943).

[17] P. 12 of ftn. 15 (1917).

[18] Pp. 151–52 of ftn. 1 (1943).

indices determined range from 17 to 98, with fairly even distribution between these extremes. As the sand is quite well sorted, in general, the many high indices are not surprising.

RIPPLE ORIENTATION

A distinctive feature of ripple marks in the Coconino sandstone is that the crests and troughs extend up and down the lamination slopes, paralleling the direction of dip. Without exception, all specimens found *in situ* have shown such orientation. This characteristic is consistent with the theory that they were developed among lee-side deposits along the wings of barchan dunes.

Observations of modern barchans in the Navajo Indian Reservation indicate that gentle winds often move across the steep lee faces on the dune wings, parallel to the direction of extension, re-working the sand surfaces and developing ripple marks (Fig. 1). Under these conditions, ripple marks invariably strike in the direction of sand slope which is at right angles to that of wind movement. In contrast, when the wind approaches from other directions, rippleless laminae develop for the air currents then are diverted by the dune, causing sand accumulation by the normal sliding or avalanching method, as opposed to the saltation common to windward-side deposits.

Ripple marks on the windward sides of modern dunes show no constancy of orientation with respect to slope comparable to that found on barchan wings of the lee sides. Low-angle approach surfaces normally are covered with ripples in which crests and troughs strike in many directions and have no apparent relation to slope. Many crests show branching, curving, or irregular patterns.

Ripple marks formed by stream currents seldom, if ever, develop parallel to the direction of slope on steeply dipping beds. Movement of water downgrade must ordinarily cause any ripples that form to be at right angles to the direction of inclination.

Wave ripples, in contrast to those of water currents, have their orientation controlled by the direction of wave movement, independently of the configuration of the bottom. O. F. Evans[19] describes waves coming in toward the shore of a lake and moving nearly parallel to the axis of the subaqueous ridges, thus forming asymmetrical ripple marks with crests and troughs parallel to the direction of dip. Such ripple marks may resemble in attitude those of lee-side dune ripples, but their position relative to the sand slope will be a matter of coincidence varying with the relief of the sea floor, whereas the wind ripples are formed exclusively with the one attitude.

EXTENT AND PARALLELISM

Crests and troughs of ripple marks in the Coconino sandstone show a high degree of parallelism and a marked constancy of direction (Fig. 3). Although the maximum extent of most rippled surfaces in this formation is not determinable because of incomplete exposures, it is clear from available outcrops that ripple marks vary but little from top to bottom of the steeply dipping surfaces on which they are formed.

A regular, even, ripple pattern, such as that found in the Coconino sandstone, is a common feature of steeply sloping, lee-side deposits on modern barchan wings (Figs. 1 and 2) and, to a less extent, between the wings. It is readily explained as the result of ripple-forming

[19] "The Relation between the Size of Wave-formed Ripple Marks, Depth of Water, and the Size of the Generating Waves," *Jour. Sed. Petrol.*, Vol. XII (1942), p. 34.

winds which are able to approach such surfaces from one direction only. In contrast, ripple marks on the windward sides of dunes, where winds constantly shift direction, develop many bifurcations, variations in trend, and changes in size.

Many wave-formed ripple marks develop with a high degree of parallelism and straightness of crest and trough. Such ripple marks form normal to the wave direction, irrespective of bottom topography and, therefore, are constant over wide areas. On the other hand, ripple marks formed by stream or other currents normally reflect great variation in direction and speed, and so they have extremely irregular patterns when considered over a large surface.

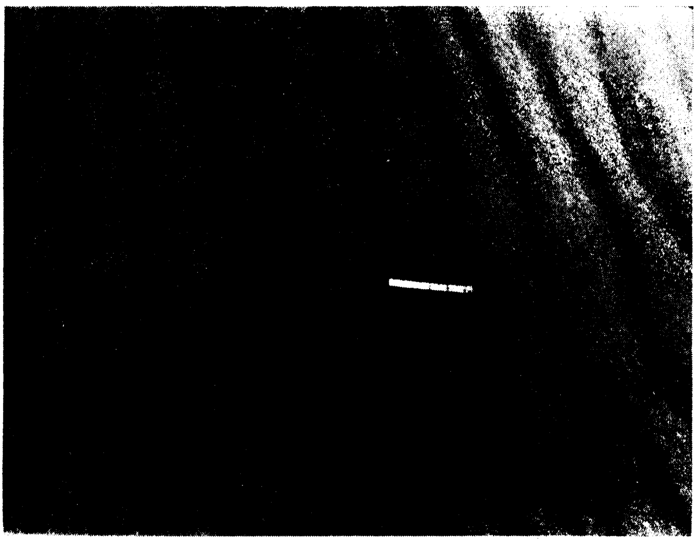


FIG. 2.—Stabilized wind-ripple marks on lee side of barchan dune near Tuba, Arizona

ROUNDED CRESTS

Ripple marks in the Coconino sandstone show a definite rounding of crests (Fig. 3). In some, the degree of rounding is so marked that it is almost equal to that of the troughs. In others, the contrast between gentle approach slopes and steep lee slopes is more apparent; yet surface rounding is still considerable.

A probable explanation for the high degree of rounding in ripple-mark crests in the Coconino sandstone has been obtained through examination of modern dune ripples in the Navajo Indian Reservation. It has been observed that freshly formed ripple marks are only moderately well rounded, whereas those that have been allowed to settle are flattened somewhat and have assumed a considerably more rounded form (Fig. 2). This suggests that those wind ripples that are permanently preserved in the geologic record are the ones that have had opportunity to stabilize between periods of sand accumulation and that for this reason rounding of crests is pronounced.

Well-rounded crests may also be found on ripple marks formed by water. In dis-

cussing wave ripples, O. F. Evans[20] states that "when a heavy sea is followed by a long period of gentle wave movement, a rounding and smoothing of the ripples occur. . . ." Thus, rounded crests are not necessarily indicative of eolian origin and are found among both water and wind-formed ripples where an alteration in form is developed during a period of stabilization.

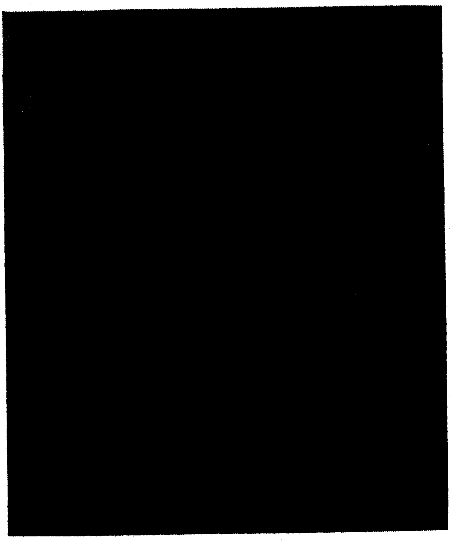


FIG. 3.—Ripple marks in Coconino sandstone near Ash Fork, Arizona.

GRAIN SORTING

Slabs of Coconino sandstone containing ripple marks have been cut and polished to show the distribution of sand grains in cross sections through the ripple series. In some samples no appreciable variations in grain size could be detected, the ripples apparently being formed in uniform, well-sorted sand. In others, grains of more than one grade size are present. Here the coarser grains form ripple crests, and the finer grains are in the troughs. Where rippling is absent and grains are of two grade sizes, the sand is separated into thin layers or laminae, consisting of an alternation of coarser and finer material.

The sorting of sand grains according to size in wind-ripple-mark specimens is a function of the process of deposition involved. The sand is piled up by saltation into a series of ridges in which the coarsest grains invariably collect at the crests, because, as explained by Bagnold,[21] they are less easily moved and, therefore, protect the crests, allowing them to rise into a region of stronger wind than would otherwise be possible. In contrast to this, sorting in the flat laminae, where rippling is absent, results from a different cause. Here the layers are developed by slumping or avalanching of sand grains (as opposed to saltation), a process in which the grains invariably become sorted according to grain size.

SLUMP MARKS

GENERAL

On the lee side of barchan or transverse dunes, especially in the steep middle portions, sand lies near the angle of repose with a surface that normally is smooth and flat and without ripple marks. This type of surface develops because saltating grains coming to rest on the upper part of the lee slope build up an oversteepened profile that is relieved periodically by avalanching. Such avalanching, as pointed out by Bagnold,[22] comes as a series of discontinuous jerks with successive miniature landslides

[20] "Effects of Change of Wave Size on the Size and Shape of Ripple Marks," *Jour. Sed. Petrol.*, Vol. XIII (1943), p. 38.

[21] P. 152 of ftn. 1 (1943).

[22] *Ibid.*, pp. 203 and 239; also, "The Transport of Sand by Wind," *Geog. Jour.*, Vol. LXXXIX (1937), p. 433.

forming to leeward each time the equilibrium of the steep slope is disturbed beyond a critical value. Although mass movements of this type result largely from overloading at the top of the lee slope, locally large-scale avalanches form from the weight of an animal walking along the top or as a result of undermining by a cross-current of wind. These avalanches cause the development on sand surfaces of a series of variable and irregular lines roughly parallel to the direction of slope and marking the borders of the sand mass that has slumped downward. For such lines the term "slump mark" is proposed.

Typical slump marks are preserved on slabs of Coconino sandstone in numerous localities. An attempt has been made through experimentation to determine the factors controlling their origin and the characteristics developed under varying conditions. The project was relatively simple, as it required only an artificial dune and an opportunity to observe and record features developed through disturbance of sand resting at the angle of repose. In order to note any effects of differences in grain size, the sand was first sieved into three grades (0.5–1.0, 0.25–0.5, and 0.125–0.25 mm.), each of which was dyed a distinctive color, following Bagnold's method. They were then thoroughly mixed and allowed to become sorted by natural processes during avalanching. In order to determine differences in slump marks due to degree of dampness, experiments made in dry sand were repeated in wet sand, damp sand, and wet sand that had dried.

SLUMP MARKS IN DRY SAND

In plan view, slump marks developed in dry sand show subparallel lines passing up and down the bedding slope and converging toward an apex near each end except where limited by the upper or lower margins of the sand surface. These lines that bound the area of slumping may be in the form of raised margins or of margins that are depressed, depending on local conditions. Likewise the area bounded by them may stand up as a low platform; or it may form a shallow, flat trough. Commonly both types are found in different parts of the same mass of slumped sand, the raised platform being in the lower portion, owing to the addition of sand that has slid down from above. If the sand is of several size grades, the coarser grains tend to separate out and slide down on top, forming the surface of the raised platform; but where sand is uniform-grained, the same surface features will develop, with texture playing no part.

Slump marks in dry sand show an infinite variety of form and detail. Only the simplest, basic type, as developed experimentally, has been described above, whereas in modern dunes compound forms with many nearly parallel lines commonly develop from a series of partially superimposed avalanches (Fig. 4). In some places, groups of overlapping tongues or funnels mark the bottoms of slumped sand patches; in other places, series of tiny, rounded, steplike structures develop because of the breaking-off of sand layers along lamination planes where contrasting grain size is involved. Several different types of dry-sand slump marks have been recognized on surfaces of the Coconino sandstone (Fig. 5).

Because avalanche surfaces characteristically develop on all parts of the lee sides of barchan and transverse dunes, the associated slump lines commonly serve as initial irregularities necessary to the development of ripple marks that form on the dune wings. In brief, the slump structures of a barchan center

commonly converge into well-formed ripple marks on the dune flanks.

SLUMP MARKS IN DAMPENED SAND

When dune sand is dampened, as by mist or dew (conditions simulated in experiments by using a fine spray), a crust, varying in thickness according to the depth of moisture penetration, is developed. Subsequent slumping causes this crust to break up into a jumble of irregular, thin patches of sand. Sand curls and crusty fragments form a confused mass. Similar results are obtained after the sand has dried out. This type of slump mark has not been recognized in any specimens of Coconino sandstone.

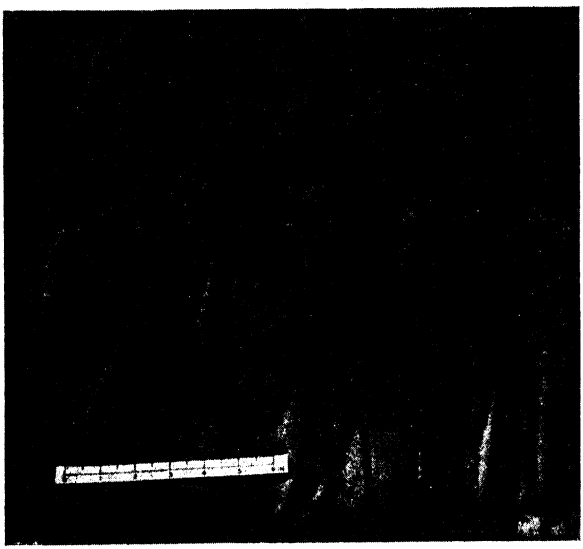


FIG. 4.—Slump marks in dry dune sand, Tuba, Arizona

SLUMP MARKS IN WETTED SAND

Dune sand that has been thoroughly wetted develops (either before or after drying out) a very characteristic type of slump mark when caused to avalanche by overweighting from above or by undermining due to crumbling away at the base. It forms a series of miniature steps, each a few millimeters high and with platforms dipping at an angle lower than that of the avalanche slope. The amount of each rise is dependent on the thickness of individual laminae, and these in turn are probably controlled primarily by the sorting process that plays such a prominent part in the forming of lee-side avalanche deposits. These structures are sharper and more pronounced than any comparable types developed in dry sand.

Wetted-sand slump marks were first reported from the Coconino sandstone by the writer[23] when he found an excellent specimen on the Grandview trail in Grand Canyon. A correct interpretation of this feature subsequently was suggested by Reiche[24] on the basis of observations that he had made on modern dunes in New Mexico. Other speci-

[23] P. 103 of ftn. 2 (1933).

[24] P. 918 of ftn. 3 (1938).

mens of Coconino sandstone illustrating wetted-sand slump marks have since been found by the writer near Seligman, Arizona (see Fig. 6).

RAIN PATCHES

The crater-like pits formed by raindrops in mud and other fine-grained materials are familiar to most geologists and have frequently been recorded in the literature. Impressions made by rain in sand, however, do not appear to have been recognized in many places in the geologic column. W. H. Twenhofel[25] re-

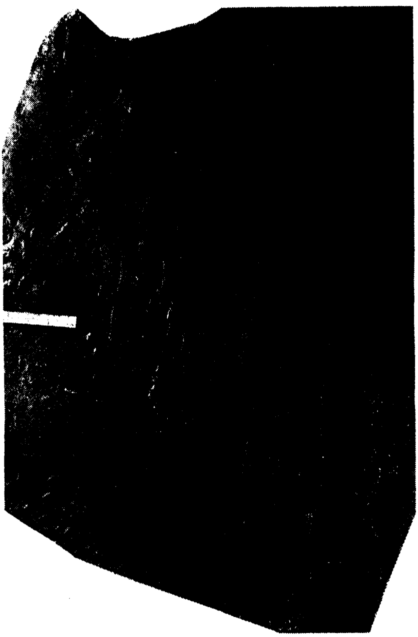


FIG. 5.—Slump marks in Coconino sandstone near Ash Fork, Arizona.

fers to them briefly, stating that "the rims are not so sharp as those bordering impressions made in mud."

Examination of rain markings in dry dune sand shows that they are different from those formed in mud. Each raindrop, as it strikes the sand, develops a thin detached layer or shell, roughly circular but with irregular margins. The larger the pits, the more accentuated are

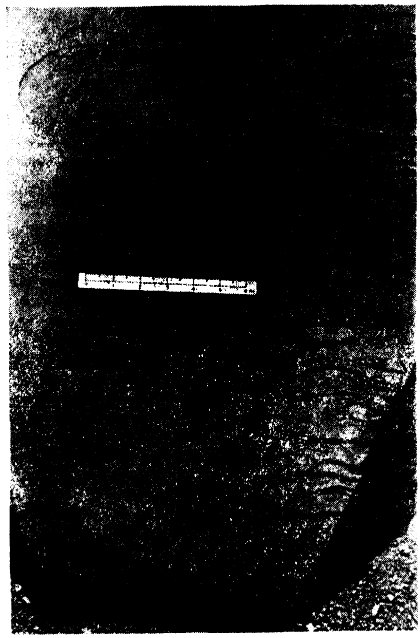


FIG. 6.—Slump marks in Coconino sandstone near Seligman, Arizona.

these irregularities. Centers of the shells are depressed, and margins are very slightly raised above the general sand-level, as with rain pits formed in mud; but the detached character of the layer of sand gives it a distinctive form (Fig. 7). After drying, the circular shells normally remain firm and retain the same characteristics as when wet.

Rain patches formed on the steep, lee-side slopes of dunes are similar to those developed on horizontal surfaces except that the shells are oriented differently with respect to the lamination plane. In

[25] *Treatise on Sedimentation* (Baltimore: Williams & Wilkins, 1926), p. 490.

general, each circular shell tends to face upward or vertically, and on the steep surface this causes the downslope rim to be raised and the upslope rim to be depressed. Furthermore, only brief showers can leave such a record, for on the surface of a dune that is thoroughly wetted the shells lose coherence. Poorly preserved markings of this type have been noted in a few places in the Coconino sandstone.

this formation, are best explained as having been formed in dry sand. The few exceptions suggest sand that had been wetted as by a rain.

Even though these surface features appear to have had their origin in dry sand, they could scarcely have been permanently preserved in such a medium. Observations on dunes and other sand deposits indicate that except possibly under special conditions of marked textural change, parting planes do not develop when the sand is dry. In contrast, where the surface has been dampened, as by a dew or fog, a firm crust is developed and remains even after drying of the surface. Such dampening by dews is a common feature in desert areas and is considered to be a likely explanation for the permanent preservation of the minor structures where additional sand is drifting over the dune crests and burying the structure-bearing layers.

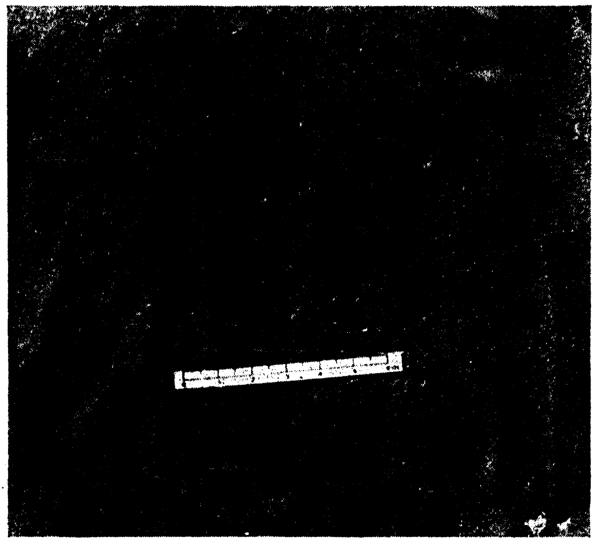


FIG. 7.—Rain patches on steeply dipping surface of dry dune sand, Tuba, Arizona

CONDITIONS FOR PRESERVING STRUCTURES

Evidence, largely based on experimental work,[26] indicates that fossil footprints which are locally common in the Coconino sandstone probably were formed, for the most part, in dry sand. Likewise, as shown in the preceding discussion, most of the minor structures, such as ripple marks and slump marks in this formation, are best explained as having been formed in dry sand.

[26] McKee, "Tracks That Go Uphill," *Plateau*, Vol. XVI (1944), pp. 61–72.

SUMMARY

The Coconino sandstone is believed to consist of the deposits of barchan and transverse dunes, as suggested by the character of its cross-lamination. It appears to be made up very largely but not entirely of steeply dipping lee-slope deposits, in which the process of mass sliding or avalanching was dominant. This method of accumulation accounts for the preponderance of smooth, even-surfaced laminae in the formation. Locally, deposits appear to have been re-worked, as by winds passing along the lee sides of barchan wings; and in these places deposition was by saltation, with consequent ripple-mark development. On surfaces formed exclusively through the process of mass sliding the principal markings developed are marginal lines of individual avalanches, and related features here referred to as "slump marks." Rare shell-like patches, probably formed by rain drops, and reptilian footprints, nearly all of which go up hill, are other surface features found in these deposits.

ACKNOWLEDGMENTS.—Experimental work was undertaken in the laboratories of the Museum of Northern Arizona during the summers of 1943 and 1944. The writer was ably assisted in this project by Messrs. L. F. Brady and J. I. Snow and by Miss Betty Lou Decker, all of the Museum staff. The writer wishes also to acknowledge with thanks the constructive criticisms and suggestions offered by Drs. H. S. Colton, Parry, Reiche, and O. F. Evans after going over the manuscript.

GEOLOGISTS STARRED, 1903–43: WHERE EDUCATED, AGE

STEPHEN S. VISHER
Indiana University, Bloomington, Indiana

ABSTRACT

The 272 geologists starred in the first seven editions of *American Men of Science* are listed, with date of birth and death (if dead); the place of their collegiate and doctoral training is given, arranged by institutions; trends in education and in institutional productivity are briefly discussed; and data on the age at starring are given.

INTRODUCTION

In the seven editions of *American Men of Science*[1] an attempt is made to sketch all living men and women who have made considerable contributions to the advancement of science in America. One thousand of the four thousand sketched in the first volume were selected (in 1903) as having made distinguished contributions, in the opinion of ten leaders in their science. The number of stars allotted to each science was roughly proportional to the number of recognized research workers in that science in 1903. The selection, after that of 1903, was partly by all previously starred workers in that science and partly by those recently nominated for starring.

During the period from 1903 to 1943, inclusive, 272 American geologists (including paleontologists, mineralogists, geophysicists, etc.) have been starred. The star implies either a large volume of good work or a considerable amount of especially original work. Although personality, position, and support also play a part in the voting, starring is a recognition which is less influenced by such secondary factors than is almost any other high honor received by scientists.

The star is a recognition which not only encourages and puts the starred men more fully on their mettle but attracts attention to their work. Starring also increases opportunities for further research, as grants and superior appointments are received comparatively readily by starred men. The various universities and other institutions employing starred scientists are placing increased value upon starring as a proof of individual merit and institutional strength. They not only attempt to retain and attract men already starred but hope to have other staff members not yet starred win this high honor; to this end they often increase facilities and otherwise encourage their more promising men.

Geology is advanced in numerous ways: by those who do high-class research, facilitate research, or aid in the effective presentation of research (editors, for example); by those who train future research men or who prepare the way for some sorts of geologic research (as, for example, topographers). It is also aided, of course, by those who, although their work is not especially original, nevertheless add substantial bits of geologic knowledge to the ever growing edifice.

Few can make major contributions; but, unquestionably, he who encourages those who have done well, makes their work more widely appreciated, and stimulates others to higher efforts is accomplishing something worth while. Hence, studies such as this are minor

[1] Lancaster, Pa.: Science Press, 1906–44.

contributions to the advancement of geology.

In this study five chief things are undertaken. (1) All the 272 geologists starred in any of the seven starrings are listed alphabetically (in List 1), with the year of their birth and death (if dead) and the date when starred (indicated by numeral of the edition in which starring occurred: 1 for 1903, 2 for 1909, 3 for 1921, 4 for 1927, 5 for 1932, 6 for 1937, and 7 for 1943.) (2) The place of collegiate training (the Bachelor's degree) is given by institutions (in List 2). (The numeral after the name dates the starring and also renders unnecessary, in most instances, the given name, which may be found in the alphabetical list.) (3) Table 1, of the number of college alumni classified by starrings, makes more evident periodic changes in institutional productivity. (4) Similar institutional lists and Table 2 are given for doctorates conferred. Table 3 summarizes some data as to degrees held and place of training. (5) Finally, some data are given on the age at starring.

THE LIST

The list of 272 starred geologists is made up of 110 selected in 1903 and 25–30 first starred in each of the other six editions. The 110 of 1903 are listed, in the order of their distinction then, on pages 1273–74 of the fifth edition of *American Men of Science*, which volume also lists (pp. 1262–63) the 25 first starred in 1932. The full official lists of those starred in the other editions have not been published, except that for 1943, with one omission, but have been partly supplied to the present writer by Editor Cattell. A few men who at the time of their starring were more distinguished in some other field than geology are included, provided they are also known as geologists. For example, several paleontologists were starred as zoölogists or botanists; several mineralogists, as chemists or physicists. Any starred man long connected with the United States Geological Survey or otherwise widely known for his geologic work is included. For example, F. W. Clarke (*Data of Geochemistry*) was a distinguished chemist of the Geological Survey and certainly is properly included here, although he was not strictly a geologist. (If any of this list not reported as dead have died, information concerning the date of their death will be appreciated.)

LIST 1

GEOLOGISTS STARRED, 1903–43

	Born	Died	Edition Starred
Alden, William C.	9-27-71		*3
Anderson, Charles A.	6-6-02		*7
Antevs, Ernst	11-20-88		*5
Arnold, Ralph	4-14-75		*2
Ashley, George H.	8-9-66		*2
Atwood, Wallace W.	10-1-72		*3
Bain, H. Foster	11-2-71		*1
Balk, Robert	5-31-99		*6
Barrell, Joseph	12-15-69	1919	*2
Barton, Donald C.	6-29-89	1939	*5
Bascom, Florence R.D.	7-14-62	1945	*1
Bassler, Ray S.	7-22-78		*3
Bastin, Edson S.	12-10-78		*3
Bateman, Alan M.	1-6-89		*5
Bayley, William S.	11-10-61	1943	*2
Becker, George F.	1-5-47	1919	*1
Beecher, Charles E.	10-9-56	1904	*1
Behre, Charles H., Jr.	3-16-96		*6
Berkey, Charles P.	3-25-67		*3
Berry, Edward Wilbur	2-10-75		*3
Billings, Marland P.	3-11-02		*7
Blackwelder, Eliot	6-4-80		*3
Blake, Wm. Phipps	6-1-26	1910	*1
Bowen, Norman L.	6-21-87		*4
Bowie, William	5-6-72	1940	*4
Bowman, Isaiah	12-26-78		*3
Bradley, Wilmot H.	4-4-99		*6
Branner, John C.	7-4-50	1922	*1
Bretz, J Harlan	9-2-82		*6
Bridge, Josiah	7-17-90		*7
Brigham, A. P.	6-12-55	1932	*1
Brooks, Alfred H.	7-18-71	1924	*1
Brush, George J.	12-15-31	1912	*2
Bryan, Kirk	7-22-88		*5

LIST 1—*Continued*

	Born	Died	Edition Starred
Bucher, Walter H.	3-12-88		*5
Buckley, Ernest R.	9-3-72	1912	*1
Buddington, A. F.	11-29-90		*6
Burbank, W. S.	3-30-98		*7
Butler, Bert S.	3-30-79		*4
Butts, Charles	9-18-63		*4
Buwalda, John P.	12-16-86		*4
Calvin, Samuel	2-2-40	1911	*1
Campbell, Marius R.	9-30-58	1940	*1
Capps, Stephen R.	10-15-81		*5
Case, Ermine C.	9-11-71		*3
Chamberlin, Rollin T.	10-20-81		*3
Chamberlin, T. C.	9-25-43	1928	*1
Chaney, Ralph W.	8-24-90		*5
Clark, William B.	12-16-60	1917	*1
Clarke, Frank W.	3-19-47	1931	*1
Clarke, John Mason	4-15-57	1925	*1
Cloos, Ernst	5-17-98		*6
Cooke, C. Wythe	7-20-87		*6
Cooper, G. Arthur	2-9-02		*7
Croneis, Carey G.	3-14-01		*7
Crosby, William O.	1-14-50	1925	*1
Cross, Whitman	9-1-54		*1
Cushing, H. P.	10-10-60	1921	*3
Cushman, Joseph A.	1-31-81		*5
Dall, William H.	8-21-45	1927	*1
Daly, Reginald A.	5-19-71		*2
Dana, Edward S.	11-16-49	1935	*1
Darton, Nelson H.	12-17-65		*1
Davis, William Morris	2-12-50	1934	*1
Day, Arthur L.	10-30-69		*2
De Golyer, E. L.	10-9-86		*7
Derby, Orville A.	7-23-51	1915	*1
Diller, J. S.	8-27-50	1928	*1
Dodge, Richard E.	3-30-68		*1
Dunbar, Carl O.	1-1-91		*6
Dutton, Clarence E.	5-15-41	1912	*1
Eastman, Charles R.	6-5-68	1918	*1
Eldridge, George H.	12-25-54	1905	*1
Emerson, B. K.	12-20-43	1932	*1
Emmons, Samuel F.	3-29-41	1911	*1
Emmons, William H.	2-1-76		*3
Fairchild, Herman L.	4-29-50	1943	*1
Farrington, Oliver C.	10-9-64	1933	*1
Fenneman, Nevin M.	12-26-65		*3
Fenner, Clarence N.	7-19-70		*5
Ferguson, Henry G.	6-21-82		*5
Flint, Richard Foster	3-1-02		*7
Foerste, August F.	5-7-62	1936	*5
Fontaine, William M.	21-1-35	1913	*1
Foshag, William F.	3-17-94		*6
Gale, Hoyt Stoddard	12-9-76		*3
Gannett, Henry	8-24-46	1914	*1
Gardner, Julia	82		*1
Gilbert, Grove K.	5-6-43	1918	*1
Gilluly, James	6-24-96		*6
Gilmore, Charles W.	3-11-74		*5
Girty, George H.	12-30-69	1939	*3
Goldthwait, James W.	3-22-80		*4
Grabau, Amadeus W.	1-9-70		*1
Grant, U. S.	2-14-67	1932	*1
Graton, Louis C.	6-10-80		*3
Gregory, Herbert E.	10-16-69		*2
Gregory, William K.	5-19-76		*3
Griggs, D. T.	10-6-11		*7
Grout, Frank F.	1-24-80		*4
Gruner, John W.	7-12-90		*7
Gutenberg, Beno	6-4-89		*6
Hague, Arnold	12-3-40	1917	*1
Hall, Christopher W.	2-28-45	1911	*1
Harris, Gilbert D.	10-2-64		*1
Hatcher, John B.	10-11-61	1904	*1
Haworth, Erasmus	4-17-55	1932	*1
Hay, Oliver P.	5-22-46	1930	*2
Hayes, C. Willard	10-8-59	1916	*1
Heald, K. C.	3-14-88		*4
Heilprin, Angelo	3-31-53	1907	*1
Hess, Frank L.	9-4-71		*5
Hewett, Donnel F.	6-24-81		*4
Hilgard, Eugene W.	1-5-33	1916	*2
Hill, Robert T.	8-11-58	1941	*1
Hitchcock, Chas. H.	8-23-36	1919	*1
Hobbs, William H.	7-2-64		*1
Holland, William J.	8-16-48	1932	*1
Hollick, Arthur	2-6-57	1933	*1
Holmes, Joseph A.	11-23-59	1915	*1
Hotchkiss, William O.	9-17-78		*7
Hovey, Edmund O.	9-15-62	1924	*3
Huntington, Ellsworth	9-16-76		*3
Iddings, J. P.	1-21-57	1921	*1
Ingerson, F. Earl	10-28-06		*7
Irving, John D.	8-18-74	1918	*2
Jackson, Robert T.	7-13-61		*1
Jaggar, Thomas A.	1-24-71		*2
Johannsen, Albert	12-3-71		*3
Johnson, Douglas W.	11-30-78	1944	*3
Julien, Alexis A.	2-13-40	1919	*1
Kay, George F.	9-14-73	1943	*5
Kay, G. Marshall	11-10-04		*7
Keith, Arthur	9-30-64	1944	*1
Kemp, James F.	8-14-59	1926	*1
Kerr, Paul F.	1-12-97		*7
Keyes, Charles R.	12-24-64	1942	*1
Kindle, Edward M.	3-10-69	1940	*3
King, Philip B.	9-24-03		*7
Knopf, Adolph	12-2-82		*3
Knopf, Eleanora B. (Mrs. A.)	7-15-83		*6
Knowlton, Frank H.	9-2-60	1926	*1
Krumbein, W. C.	1-28-02		*7
Kummel, Henry B.	5-25-67		*1
Lahee, Frederic H.	7-27-84		*5
Lane, Alfred C.	1-29-63		*1
Larsen, Esper	3-14-79		*4
Lawson, Andrew C.	7-25-61		*1
Lee, Willis T.	12-24-64	1926	*3
Leighton, Morris M.	8-4-87		*5
Leith, Charles K.	1-20-75		*2

	Born	Died	Edition Starred
Leverett, Frank	3-10-59	1943	*1
Levorsen, A. I.	7-5-94		*7
Lindgren, Waldemar	2-14-60	1939	*1
Longwell, Chester R.	10-15-87		*5
Loomis, F. B.	11-22-73	1937	*4
Louderback, George D.	4-6-74		*3
Loughlin, Gerald F.	12-11-80		*4
Lovering, Thomas S.	5-12-96		*6
Lull, Richard Swann	11-6-67		*3
Macelwane, James B.	9-28-83		*6
McGee, W J	4-17-53	1912	*1
McLaughlin, Donald H.	12-15-91		*6
Mansfield, George R.	8-30-75		*4
Martin, Lawrence	2-14-80		*3
Mather, Kirtley F.	2-13-88		*4
Mathews, Edward B.	8-16-69	1944	*2
Mathes, Francois E.	3-16-74		*3
Matthew, W. D.	2-19-71	1930	*3
Mead, Warren Judson	8-5-83		*4
Meinzer, Oscar E.	11-28-76		*4
Mendenhall, Walter C.	2-20-71		*2
Merriam, John C.	10-20-69		*2
Merrill, Fred J. H.	4-30-61	1916	*1
Merrill, George P.	5-31-54	1929	*1
Mertie, J. B.	1-22-88		*7
Merwin, Herbert E.	2-20-78		*5
Miser, Hugh D.	12-18-84		*4
Moore, Raymond C.	2-20-92		*5
Newell, Frederick H.	3-5-62	1932	*1
Newhouse, W. W.	12-13-97		*7
Niles, William H.	5-18-38	1910	*1
Nolan, Thomas B.	5-21-01		*6
Osborn, Henry Fairfield	8-8-57	1935	*1
Palache, Charles	7-18-69		*2
Penfield, Samuel L.	1-16-56	1906	*1
Penrose, Richard A. F., Jr.	12-17-63	1931	*1
Pirsson, L. V.	11-3-60	1919	*1
Powers, Sidney	9-10-90	1932	*4
Prosser, Charles S.	3-24-60	1916	*1
Pumpelly, Raphael	9-8-37	1923	*1
Ransome, Frederick L.	12-2-68	1935	*1
Raymond, Percy Edward	5-30-79		*4
Reed, Ralph D.	4-21-89	1940	*6
Reeside, John B.	6-24-89		*5
Reid, Harry Fielding	5-18-59	1944	*1
Rice, William North	11-21-45	1928	*1
Rich, John L.	12-1-84		*6
Ries, Heinrich	4-30-71		*1
Rogers, Austin F.	8-15-77		*4
Ross, Clarence S.	9-20-80		*5
Rubey, William W.	12-19-98		*6
Ruedemann, Rudolph	10-16-64		*3
Russell, Israel C.	12-10-52	1906	*1
Safford, James M.	8-13-22	1907	*1
Sales, Reno H.	9-10-76		*7
Salisbury, Rollin D	8-17-58	1922	*1
Schairer, J. F.	4-13-04		*7
Schaller, Waldemar T.	8-3-82		*3
Schuchert, Charles	7-3-58	1942	*1
Scott, William B.	2-12-58		*1
Sellards, Elias H.	5-2-75		*5
Shaler, Nathaniel S.	2-21-41	1906	*1
Shand, S. James	10-29-82		*7
Shaw, Eugene W.	7-29-81	1935	*3
Shepard, Francis P.	5-10-97		*6
Short, M. N.	3-21-89		*7
Simpson, George G.	6-11-02		*7
Singewald, Joseph T., Jr.	9-25-84		*4
Smith, Eugene A.	10-27-41	1927	*1
Smith, George Otis	2-22-71	1944	*2
Smith, Jas. Perrin	11-27-64	1931	*1
Smyth, Henry L.	1-11-62	1944	*1
Spencer, Arthur Coe	9-27-71		*2
Spencer, J. W.	3-26-51	1921	*2
Springer, Frank	6-17-48	1927	*4
Spurr, Josiah E.	10-1-70		*1
Stanton, Timothy W.	9-21-60		*1
Stephenson, Lloyd W.	8-31-76		*4
Stevenson, John J.	10-10-41	1924	*1
Stock, Chester	1-28-92		*6
Stose, George W.	10-5-69		*4
Tarr, Ralph S.	1-15-64	1912	*1
Taylor, Frank B.	11-23-60	1938	*1
Thom, William Taylor, Jr.	6-9-91		*5
Todd, James E.	2-11-46	1922	*1
Tolman, Cyrus F.	6-2-73	1942	*3
Trask, Parker D.	5-7-99		*6
Trowbridge, Arthur C.	3-4-85		*6
Tunell, George	4-4-00		*7
Turner, H. W.	8-22-57	1937	*1
Twenhofel, William H.	4-16-75		*4
Ulrich, Edward O.	2-1-57	1944	*2
Upham, Warren	3-8-50	1934	*1
Van Hise, C. R.	5-29-57	1918	*1
Vaughan, T. Wayland	9-20-70		*1
Wadsworth, M. E.	5-6-47	1921	*1
Walcott, Charles D.	3-31-50	1927	*1
Ward, Lester Frank	6-18-41	1913	*1
Washington, Henry S.	1-15-67	1934	*1
Waters, A. C.	5-6-05		*7
Watson, Thomas L.	9-5-71	1924	*3
Weaver, C. E.	5-1-80		*7
Weed, Walter H.	5-1-62		*1
Weller, Stuart	12-26-70	1927	*1
Wentworth, Chester K.	5-7-91		*5
Wheeler, Homer Jay	9-2-61		*2
White, Charles A.	1-26-26	1910	*1
White, David	7-1-62	1935	*1
White, Israel C.	11-1-48	1927	*1
Whitfield, Robert P.	5-27-28	1910	*1
Wieland, George R.	1-24-65		*3
Williams, H. S.	3-6-47	1918	*1
Williams, Howel	10-12-98		*7
Willis, Bailey	5-31-57		*1
Williston, Samuel W.	7-10-52	1918	*1
Winchell, Alexander N.	3-2-74		*2
Winchell, Newton H.	12-17-39	1914	*1
Wolff, John E.	11-21-57	1940	*1

LIST 1—*Continued*

	Born	Died	Edition Starred
Woodring, Wendell P.	6-13-91		*5
Woodward, Robert S.	7-21-49	1924	*1
Woodworth, Jay B.	1-2-65	1925	*1
Wrather, William E.	1-20-83		*6
Wright, Frederick E.	10-16-77		*2

COLLEGIATE TRAINING

Grouped by undergraduate education, the starred geologists are found to have graduated from 81 American colleges or universities. Seven institutions graduated 9 or more—a total of 118, or nearly one-half of the total having American college degrees (Harvard, 38; Yale, 22; Cornell, 16; California, 12; Chicago, 12; Amherst, 9; and Columbia, 9). Eight institutions graduated from 3 to 7; 19 graduated 2 each; and 46 graduated 1 each (see List 2 and Table 1).

In the collegiate training of the older group (starred in 1903) Harvard, Yale, Cornell, and Amherst were pre-eminent, with 7–21 to each. Harvard also led in the 1932 group and tied for first place for the 1943 group. Yale ranked first for those starred in 1909 and tied for first for the 1943 group. Chicago ranked first for the group starred in 1921, tied for first for the 1937 and 1943 groups, and stood second for 1932. California tied for second place for the 1921 group and for first in the 1937 and 1943 groups. Cornell stood third in 1903 and tied for second in 1921 and for first in 1927. For the last three starrings combined, Chicago and Harvard tied for first; California and Yale for third; Denison, Hopkins, and Iowa for fifth. Of the 248 starred geologists who graduated from American colleges, many more than half had as an undergraduate at least one starred geologist as a teacher. It appears that most subsequently starred geologists chose their profession before graduating from college.

The most important influence in the

TABLE 1*

COLLEGIATE DEGREES TO STARRED GEOLOGISTS

Institution	Year in Which Starred								
	1903	1909	1921	1927	1932	1937	1943	Total	1932–43
Amherst	7	0	0	2	0	0	0	9	0
Beloit	3	0	1	1	0	0	0	5	0
California	1	1	3	2	0	3	2	12	5
Chicago	0	0	4	0	3	3	2	12	8
Columbia	5	1	2	0	1	0	0	9	1
Cornell	9	1	3	2	0	1	0	16	1
Denison	0	0	0	1	2	0	1	4	3
Harvard	21	1	2	1	5	1	2	38	8
Hopkins	1	1	0	1	1	1	1	6	3
Iowa	1	0	0	0	1	0	2	4	3
Kansas	1	0	1	1	1	1	0	5	2
Lehigh	0	1	0	2	0	0	0	3	0
Massachusetts Institute of Technology	3	0	1	2	0	0	1	7	1
Minnesota	1	1	1	1	0	1	1	6	2
Wisconsin	3	1	0	1	0	0	1	6	1
Yale	11	3	2	1	1	2	2	22	5

* Also, 2 each at Bryn Mawr, Cincinnati, Colby, Colgate, Cornell (Iowa), Illinois, Iowa State College, Maine, Massachusetts College, Michigan, Middlebury, Missouri, New Mexico, New York University, Oberlin, Princeton, Queen's, Stanford, Toronto, Washington (Seattle), and Wesleyan.

training of leaders in research clearly appears to be enthusiastic teachers of their science. Opportunities for employment undoubtedly are also highly significant, as are the ready availability of unsolved problems which are not too difficult for beginners. Without such problems, inspiring teachers have serious difficulty in developing productive followers. Without well-qualified teachers but with not too difficult research problems available, some scientists develop, largely self-taught. Nevertheless, even of the older starred geologists, each of a vast majority has named, on occasion, an enthusiastic teacher of geology as deserving great credit for having deeply interested him in the subject and started him on his scholarly career and for having provided vital parts of the background of training, objectives, and self-confidence which enabled him to attain a high level of research achievement. Often, of course, more than one teacher contributed notably to the inspiration of the embryo geologist.

Former teachers, now dead, who had two or more of their special undergraduate students later win stars include: Bascom (Bryn Mawr), Bayley (Colby, Illinois), Carney (Denison), T. C. Chamberlin (Beloit, Chicago), J. D. Dana (Yale), Davis (Harvard), Emerson (Amherst), Haworth (Kansas), Salisbury (Chicago), Shaler (Harvard), Tarr (Cornell), and H.S. Williams (Cornell, Yale). (Additions to this list of living men in this category, as well as deceased ones, will be welcomed.)

Although many colleges have given undergraduate work to starred geologists, only 3 were graduated from one of the numerous agricultural and engineering schools, and apparently none from the many state teachers or Catholic colleges.

LIST 2

COLLEGES FROM WHICH STARRED GEOLOGISTS RECEIVED THE BACHELOR'S DEGREE

AMHERST: Clark, 1; Clarke, J. M., 1; Cross, 1; Emerson, 1; Hitchcock, 1; Holland 1; Kemp, 1; Loomis, 4; Mansfield, 4

BELOIT: Chamberlin, T. C., 1; Huntington, 3; Kummel, 1; Meinzer, 4; Salisbury 1

BRYN MAWR: Gardner, 6; Mrs. Knopf, 6

CALIFORNIA: Buwalda, 4; Foshag, 6; Knopf, A., 3; Larsen, 4; Louderback, 3; McLaughlin, 6; Palache, 2; Ransome, 1; Schaller, 3; Short, 7; Stock, 6; Weaver, 7

CHICAGO: Atwood, 3; Behre, 6; Blackwelder, 3; Capps, 5; Chamberlin, R. T., 3; Chaney, 5; Flint, 7; Krumbein, 7; Tolman, 3; Trowbridge, 6; Wentworth, 5; Wrather, 6

CINCINNATI: Bassler, 3; Bridge, 7

COLBY: Mathews, 2; Smith, G. O., 2

COLGATE: Brigham, 1; Cooper, 7

COLUMBIA: Gregory, W., 3; Fenner, 5; Hollick, 1; Irving, 2; Matthew, 3; Merrill, 1; Ries, 1; Weed, 1; Willis, 1

CORNELL: Ashley, 2; Branner, 1; Butler, 4; Cushing, 3; Derby, 1; Fairchild, 1; Graton, 3; Harris, 1; Hill, 1; Holmes, 1; Prosser, 1; Raymond, 4; Rich, 6; Martin, 3; Weller, 1; White, D., 1

CORNELL COLLEGE (Iowa): Alden, 3; Calvin, 1

DENISON: Croneis, 7; Foerste, 5; Mather, 4; Moore, 5

HARVARD: Barton, 5; Becker, 1; Billings, 7; Bowman, 3; Brooks, 1; Clarke, F. W., 1; Cushman, 5; Davis, 1; Diller, 1; Dodge, 1; Eastman, 1; Eldridge, 1; Emmons, 1; Ferguson, 5; Gale, 3; Gannett, 1; Goldthwait, 4; Jackson, 1; Jaggar, 2; Keith, 1; Lahee, 5; Lane, 1; Merwin, 5; Niles, 1; Penrose, 1; Shaler, 1; Shepard, 6; Smyth, 1; Spurr, 1; Tarr, 1; Tunell, 7; Vaughan, 1; Woodworth, 1; Wolff, 1

JOHNS HOPKINS: Bayley, 2; Cooke, 6; Mertie, 7; Reeside, 5; Reid, 1; Singewald, 4

ILLINOIS: Johannsen, 3; Ross, 5

IOWA: Kay, M., 7; Keyes, 1; King, 7; Leighton, 5

IOWA COLLEGE: Leverett, 1; Springer, 4

KANSAS: Case, 3; Dunbar, 6; Haworth, 1; Rogers, 4; Sellards, 5

LEHIGH: Barrell, 2; Bowie, 4; Hewett, 4

MAINE: Farrington, 1; Merrill, 1

MASSACHUSETTS INSTITUTE OF TECHNOLOGY: Burbank, 7; Crosby, 1; Grabau, 1; Loughlin, 4; Matthes, 3; Newell, 1; Stose, 4
MICHIGAN: Bastin, 3; Winchell, N., 1
MIDDLEBURY: Knowlton, 1; Hall, 1
MINNESOTA: Berkey, 3; Grant, 1; Grout, 4; Levorsen, 7; Lovering, 6; Winchell, A. N., 2
MISSOURI: Rubey, 6; Longwell, 5
NEW MEXICO: Bryan, 5; Johnson, 3
NEW YORK UNIVERSITY: Russell, 1; Stevenson, 1
OBERLIN: Hayes, 1; Todd, 1
PRINCETON: Osborn, 1; Scott, 1
QUEEN'S (Ont.): Bateman, 5; Bowen, 4
STANFORD: Arnold, 2; Hess, 5
TORONTO: Kay, G., 5; Lawson, 1
WASHINGTON (Seattle): Gilluly, 6; Waters, 7
WESLEYAN (Conn.): Rice, 1; Lee, 3
WISCONSIN: Bascom, 1; Buckley, 1; Hotchkiss, 7; Leith, 2; Mead, 4; Van Hise, 1
YALE: Blake, 1; Bradley, 6; Brush, 2; Bryan, 5; Dana, 1; Day, 2; Dutton, 1; Girty, 3; Gregory, H. E., 2; Hague, 1; Hatcher, 1; Hovey, 3; Iddings, 1; Nolan, 6; Penfield, 1; Pirsson, 1; Schairer, 7; Simpson, 7; Twenhofel, 4; Washington, 1; Williams, H. S., 1; Williston, 1

INSTITUTIONS GRADUATING ONE EACH

ALABAMA: Smith, E. A., 1; ALBION: Bretz, 6; ALBRIGHT: Woodring, 5; ALFRED: Butts, 4; ARKANSAS: Miser, 4; BOWDOIN: Wadsworth, 1; BROWN: Buddington, 6; CASE: Spencer, A. C., 2; CENTRAL (Mo.): Emmons, W., 3; COLORADO: Stanton, 1; COLORADO COLLEGE: Heald, 4; DARTMOUTH: Upham, 1; EUREKA (Ill.): Hay, 2; GEORGE WASHINGTON: Ward, 1; HEIDELBERG (Ohio): Fenneman, 3; HIRAM: Reed, 6; INDIANA: Kindle, 3; KANSAS STATE: Williston, 1; LENOX COLLEGE (Iowa): Merriam, 2; MASSACHUSETTS COLLEGE: Wheeler, 2; MOORES HILL (Ind.): Bain, 1; OCCIDENTAL: Kerr, 7; OHIO: Griggs, 7; OHIO (Athens): Safford, 1; OHIO NORTHERN: Mendenhall, 2; OHIO WESLEYAN: Shaw, 3; OKLAHOMA: De Golyer, 7; PENNSYLVANIA STATE COLLEGE: Wieland, 3; POMONA: Anderson, 7; ROCHESTER: Gilbert, 1; RUTGERS: Lull, 3; SCIO COLLEGE (Ohio): Stephenson, 4; SIMMONS: Ingerson, 7; ST. LOUIS: Macelwane, 6; TEXAS: Trask, 6; TRINITY: Bowie, 4; TULANE: Vaughan, 1; UNION: Julien, 1; VIRGINIA: Fontaine, 1; VIRGINIA STATE COLLEGE: Watson, 3; WASHINGTON AND LEE: Thom, 5; WEST VIRGINIA: White, I. C., 1; WILLIAMS; Powers, 4; WOFFORD (S.C.): Smith, J. P., 1; WORCESTER: Hobbs, 1; WYOMING: Gilmore, 5

DOCTORAL TRAINING

For their graduate work (List 3 and Table 2) most starred geologists attended relatively few universities. Very few took their doctorate at an American university which did not have two or more starred geologist professors. This implies that, with few exceptions, expert training is needed. Of the 1903 and 1909 groups, 14 obtained doctorates from Europe; but of the younger native geologists, only 1 (Bucher) did so. In recent years America has afforded at least as good advanced geological training as Europe, as well as more opportunities for well-qualified geologists. This helps to explain the immigration here of a number of capable geologists (6 among those starred during 1932–43).

Of the 188 starred geologists with regular doctorates (not honorary degrees), 102 received their degrees from four American universities (Yale 37, Chicago 23, Harvard 22, or Hopkins 20); and 24 received them in Europe. Of the remaining 62, 12 received their doctorate at California and 10 at Columbia. (For the complete list see List 3.) Of the recent starrings (1932–43), Yale conferred doctorates upon 15, Chicago 11, Harvard 11, California 6, Hopkins 6; no other university on more than 2 (Table 2).

Table 2 discloses interesting variations from time to time in institutional doctoral productivity. California stood low, or rather low, until 1937, when it tied for first place. Chicago led in 1921 and tied for first place in 1932 and 1937. Columbia and Cornell tied for third place in 1921 but stood low at the other starrings. Harvard stood high in 1903, 1927, 1932, and 1943. Hopkins stood

second in 1903 and fourth in 1932 and 1937. Yale led in 1903, 1927, 1943, and tied in first in 1932 and 1937.

The fluctuation in output of subsequently starred men correlates partly with the age of the leading teachers. Elderly men, despite perhaps earnest efforts strongly to stimulate their students to high achievement, are usually distinctly less successful in doing so than are men less removed from their students in age and achievement. When the gap in age and scholarly achievement between professor and student is large, the close fellowship which apparently is almost essential for effective stimulation can seldom obtain. Thus an enthusiastic, able instructor usually has far more influence on his students than the same individual has twenty or thirty years later, after he has become, perhaps, a world-renowned geologist.

Some mention of graduate work which did not lead to a doctorate is appropriate. A considerable number of starred geologists did not acquire an earned doctorate—55 per cent of those starred in 1903, about one-fifth of those of the next three starrings, and about one-tenth of those of the last three. Many of the older geologists had graduate work (without receiving a doctorate) in Germany or at Harvard. Of those starred during 1921–43 without a doctorate, 3 had graduate work at Chicago (Shaw, Tolman, Wrather), 2 at Yale (Heald, Rubey), 2 at Harvard (Gale, Matthes), and 1 each at Columbia (Sales), Cornell (Butler), and Massachusetts Institute of Technology (Burbank).

TABLE 2*

DOCTORATES TO GEOLOGISTS—SUMMARY

Institution	Year in Which Starred								
	1903	1909	1921	1927	1932	1937	1943	Total	1932–43
California	1	1	2	2	0	4	2	12	6
Chicago	2	0	8	2	5	4	2	23	11
Columbia	2	1	4	1	1	0	1	10	2
Cornell	1	0	4	0	0	1	0	6	1
George Washington	3	0	1	0	0	0	0	4	0
Harvard	6	2	0	3	5	1	5	22	11
Hopkins	7	4	1	2	3	2	1	20	6
Minnesota	0	0	1	0	0	1	1	3	2
Princeton	1	0	0	0	0	1	1	3	2
Stanford	0	2	0	0	0	1	1	4	2
Wisconsin	2	1	0	1	0	0	1	5	1
Yale	8	3	6	5	5	4	6	37	15
Foreign universities	8	6	2	1	2	3	2	24	7

* Also, 2 each at Massachusetts Institute of Technology (4, 7), New York University (1); 1 each at Bryn Mawr (6); Illinois (5); Iowa (5).

MOBILITY

There is a correlation between the number born in a region, state, or city and the number who had their college and graduate work there. However, detailed studies prove that starred geologists are highly mobile. Many did not attend the nearest good school; most of them went some hundreds of miles; and a few crossed the continent for their Bachelor's degrees, as did several for their doctorates.

A study of the approximate distance

from their birthplace to the college from which they graduated has been made for 205 geologists, with the following results: 5 native-American starred geologists graduated from a college more than 2,000 miles from their birthplace; 17 graduated from one 1,000–2,000 miles away; 30 from one 500–1,000 miles away; 34 from one 250–500 miles away; 46 from one 100–250 miles away; 47 from one 25–100 miles distant; and 26 from one within 25 miles of their birthplace.

For their graduate work, including formal study in European universities, 32 native-American starred geologists went more than 3,000 miles from their birthplace; 23 went 1,000–3,000 miles; 39 went 500–1,000 miles; 29 went 250–500 miles; 21 went 100–250 miles; 19 went 25–100 miles; while 15 received their graduate schooling within 25 miles of their birthplace. These latter were largely persons born in or near Boston, New York, Chicago, Baltimore, or San Francisco.

Despite the fact that the average starred geologist attended a college about 200 miles from his birthplace and had graduate work in a university more than 500 miles from his birthplace, nevertheless the large sections of the nation which were the birthplace of many subsequently starred geologists likewise trained many.

TABLE 3

SOME TRENDS BASED ON DECADE OF BIRTH

	Born before 1840*	Decade in Which Born						
		1840's*	1850's	1860's	1870's	1880's	1890's	1900's
Total number	25	27	39	54	54	43	27	14
No college degree	5	2	6	3	1	0	0	0
No graduate work	19	8	14	7	6	2	0	0
No doctorate	22	16	25	15	12	4	3	0
College degree from small college	8	13	8	14	16	9	5	4
College degree from a university	11	12	19	27	23	20	12	10
Chicago	0	0	0	0	1	4	3	2
Harvard	0	5	3	8	5	5	1	2
Hopkins	0	0	0	2	0	4	0	0
Yale	6	4	1	7	1	0	2	2
California	0	0	0	2	2	5	3	0
Other state universities	0	1	6	8	13	11	11	4
Doctorate from Europe	1	3	4	7	2	5	2	0
Doctorate from state universities	0	2	3	3	5	6	6	1
Doctorate from endowed universities	2	6	7	28	34	27	17	13
Chicago	0	0	0	2	5	7	4	2
Harvard	0	1	1	4	7	6	2	4
Hopkins	0	0	2	8	4	5	2	0
Yale	2	3	4	8	6	6	4	6

* Seventeen distinguished geologists who died before starring was done are included in this table. Seven were presidents of the Geological Society of America. Doubtless, all would have been starred had they lived to 1903. The 17 are: E. D. Cope, J. D. Dana, G. Dawson, J. Dawson, Jas. Hall, F. V. Hayden, T. S. Hunt, A. Hyatt, Jos. LeConte, J. Leidy, J. P. Lesley, O. C. Marsh, J. S. Newberry, Edw. Orton, J. W. Powell, G. H. Williams, and Alex Winchell.

SOME EDUCATIONAL TRENDS

Table 3 reveals some trends as to place and amount of schooling. A trend clearly evident is for more general academic training. Of the older starred geologists, a tenth were not college graduates; relatively few had doctorates. Of

those born since 1870, only 1 is not a college graduate, and only 7 born since 1879 do not have the doctorate.

Another trend is toward the increasing importance of the state universities in the training of geologists and the relative decline for the endowed colleges and universities, especially those of the Northeast. However, of the 14 born since 1899, only 4 got their college degrees from state universities (2 from Iowa), and only 1 his doctorate.

The declining importance of Europe in the doctoral training of starred geologists is notable. In brief, of those born before 1860 who had earned doctorates, Germany conferred them upon about one-third; of those born in the sixties, about one-sixth; of those born in the seventies, about one-twentieth. Of the native-American starred geologists born since 1888, not one received a doctorate in Europe.

LIST 3

UNIVERSITIES FROM WHICH STARRED GEOLOGISTS RECEIVED DOCTORATE

BRYN MAWR: Knopf, Mrs. A., 6;
CALIFORNIA: Anderson, 7; Buwalda, 4; Foshag, 6; Knopf, A., 3; Larsen, 4; Louderback, 3; Macelwane, 6; Palache, 2; Stock, 6; Trask, 6; Weaver, 7; Ransome, 1
CHICAGO: Alden, 3; Atwood, 3; Bain, 1; Bastin, 3; Behre, 6; Blackwelder, 3; Bretz, 6; Capps, 5; Case, 3; Chamberlin, R. T., 3; Chaney, 5; Emmons, 3; Fenneman, 3; Flint, 7; Kay, G., 5; Krumbein, 7; Kummel, 1; Leighton, 5; Mather, 4; Meinzer, 4; Moore, 5; Shepard, 6; Trowbridge, 6
COLUMBIA: Fenner, 5; Gregory, W., 3; Irving, 2; Johnson, 3; Kay, M., 7; Lull, 3; Matthew, 3; Merrill, 1; Ries, 1; Rogers, 4
CORNELL: Cushing, 3; Graton, 3; Martin, 3; Prosser, 1; Rich, 6; Watson, 3
GEORGE WASHINGTON: Bassler, 3; Hollick, 1; Knowlton, 1; Stanton, 1
HARVARD: Barton, 5; Billings, 7; Croneis, 7; Cushman, 5; Daly, 2; Foerste, 5; Goldthwait, 4; Griggs, 7; Jackson, 1; Jagger, 2; Lahee, 5; Lane, 1; McLaughlin, 6; Mansfield, 4; Merwin, 5; Penrose, 1; Powers, 4; Short, 7; Tunell, 7; Vaughan, 1; Wadsworth, 1; Wolff, 1
ILLINOIS: Ross, 5
IOWA: Wentworth, 5
JOHNS HOPKINS: Bascom, 1; Bayley, 2; Cooke, 6; Gardner, 6; Grant, 1; Haworth, 1; Hayes, 1; Hobbs, 1; Keyes, 1; Lee, 3; Mathews, 2; Mertie, 7; Reeside, 5; Reid, 1; Singewald, 4; Smith, G. O., 2; Spencer, A. C., 2; Stephenson, 4; Thom, 5; Woodring, 5
MASSACHUSETTS INSTITUTE OF TECHNOLOGY: Bowen, 4; Newhouse, 7
MINNESOTA: Berkey, 3; Gruner, 7; Lovering, 6
NEW YORK UNIVERSITY: Julien, 1; Stevenson, 1
PRINCETON: Bridge, 7; Buddington, 6; Osborn, 1
STANFORD: Arnold, 2; Ashley, 2; Kerr, 7; Reed, 6
WISCONSIN: Buckley, 1; Hotchkiss, 7; Leith, 2; Mead, 4; Van Hise, 1
YALE: Barrell, 2; Bateman, 5; Beecher, 1; Bowman, 3; Bradley, 6; Bryan, 5; Cooper, 7; Dana, 1; Day, 2; Dunbar, 6; Farrington, 1; Ferguson, 5; Gilluly, 6; Girty, 3; Gregory, H. E., 2; Grout, 4; Hewett, 4; Hovey, 3; Huntington, 3; Ingerson, 7; Kindle, 3; King, 7; Longwell, 5; Loughlin, 4; Nolan, 6; Raymond, 4; Rice, 1; Safford, 1; Schairer, 7; Sellards, 5; Simpson, 7; Twenhofel, 4; Waters, 7; Weller, 1; Wieland, 3; Williams, 1; Williston, 1

EUROPEAN UNIVERSITIES: Antevs, 5; Balk, 6; Becker, 1; Bucher, 5; Cloos, 6; Cross, 1; Eastman, 1; Emerson, 1; Gutenberg, 6; Hilgard, 2; Loomis, 4; Merriam, 2; Ruedemann, 3; Schaller, 3; Scott, 1; Shand, 7; Smith, E. A., 1; Smith, J. P., 1; Spencer, J. W., 2; Ulrich, 2; Washington, 1; Wheeler, 2; Williams, 7; Wright, 2

AGE WHEN STARRED

Table 4 presents facts as to age at starring. It discloses that the median age has varied from 40 to 49; a large share of the men were 45-55 when starred; 31 were past 60; and 48 were under 40 (of whom 23 were starred in 1903, 9 in 1909). Since 1909 only 2 men under 37 have been starred; but in 1903 and 1909, 12 men under 37 were starred.

Most of the men who were starred

under 35 years were born in the 1870's. Some of these had done enough work to win a star in 1903, when 110 men were starred and stars were easiest to get. This was because about a fourth of all of the geologists who had published several research papers were starred. Because of the sharp subsequent increase in the number of men who have done geological research, there have been in recent decades many more who have done considerable research; yet the number of stars is about constant. Most of those who had by 1903 held a doctorate in geology for more than a decade were starred, as well as many without a doctorate. In 1943, on the other hand, less than a twentieth of those who had held doctorates in geology for a decade were newly starred. One man of the 1903 starring was only 31 (Buckley), and three were 32 (Bain, Brooks, Ries). In 1909 Wright was starred at 32, and two at 34. None of the geologists born in the 1880's were starred in 1909; and, because no starring took place during the first World War, they had to wait twelve years for the next starring, until 1921, with the result that the youngest born in the 1880's were starred at 39. The geologists born in the 1890's did not need to wait so long for starring, as the next starring came in 1927. However, the youngest of the 1890–99 group was starred at 37—2 at 38, and 2 at 39. Only 13 geologists born after 1900 have yet been starred, at ages 32 (Griggs), 36, 37, 38, 39, 40, 41 (6), and 42.

Three of the 8 geologists starred while under 35 years before 1943 have already been elected presidents of the Geological Society of America, and a fourth is a member of the National Academy of Sciences. The recognition implied by early starring appears to have encouraged all but a few recipients to higher endeavors. The more honor is due, therefore, to the several geologists who, passed over at a previous starring and without the encouragement that this recognition affords, have had the persistence to continue their geologic studies until long after most men quit hard work and who won their stars at 65 or beyond—one at 79, another at 78, and others at 76, 70, 67, and 65 (2).

TABLE 4

AGE WHEN STARRED

Age	Year in Which Starred						
	1903	1909	1921	1927	1932	1937	1943
Median age	46	40	47	48	49	46	43
Percentage aged 45–55	33	18	56	56	40	56	25
Percentage under 45	40	65	32	32	36	44	50
Percentage over 55	27	17	12	12	24	0	14
Number past 60	19	3	1	2	2	0	4
Maximum age	81	78	61	79	70	55	67
Next oldest	77	76	59	65	61	55	65
Number under 40	23	9	2	3	0	6	5
Minimum age	31	32	39	37	40	36	32
Next youngest	32	34	39	39	41	38	38

CLARKSTON STAGE OF THE NORTHWEST PLEISTOCENE

R. L. LUPHER
State College of Washington

ABSTRACT

An episode of proglacial aggradation, here named the "Clarkston stage," is recorded in the Lewiston Basin region on the lower Snake River. Stream gravels accumulated to a depth of more than 400 feet in the Snake River canyon and in the lower portions of most tributary canyons. One or two tributary streams were ponded by the fill of the main canyon. The deposits are characterized by considerable cementation, weathering of basalt and granitic stones, and iron oxide stain. Two-story canyons were formed by cessation of erosion beneath the fill, prolonged lowering of slopes above, and later excavation of the fill. The Clarkston stage followed earlier Pleistocene deposition, deformation, and dissection but antedated the Wisconsin stage.

INTRODUCTION

The record of the latest Pleistocene of southeastern Washington has received considerable attention during the last two decades by workers who were primarily concerned with the origin of the channeled scablands and related features. Information on the stratigraphic aspects of Pleistocene history has been largely incidental to physiographic studies, and we have no clear understanding of Pleistocene chronology as a whole in that area.

Attempts to recognize a sequence of glacial and interglacial stages in the northeastern Washington area invaded by continental ice have not been successful, and it seems that a reliable Pleistocene chronology can best be worked out in southeastern Washington, where there is a considerable record of depositional and degradational episodes, deformation, renewed outpouring of lavas, and proglacial stages. Some of these events are recorded in the Lewiston Basin region about 100 miles above the mouth of the Snake River. The Asotin stage of canyon-cutting, apparently of early Pleistocene age, has been recognized there;[1] and stream and lake deposits of the Wisconsin stage are also present. It is now possible to recognize between these two extremes a new Pleistocene stage of proglacial aggradation which is named the "Clarkston stage." The primary purpose of this paper is to describe the stage in its type area centering at Clarkston, Washington. The stage is probably recorded by some of the early stream gravels, noted by earlier writers, in the region west of the Lewiston Basin; but the writer believes that correlation and tracing out of these and other deposits can be done best after a more complete Pleistocene chronology is established in critical areas.

GENERAL GEOLOGIC SETTING

The known Clarkston deposits lie a short distance inside the eastern margin of the Columbia River basalts. In this region most of the major stream courses were established on an erosion surface of low relief prior to the Pleistocene. Early Pleistocene history is imperfectly known. The Ringold formation of south-central Washington apparently is the first sedimentary record. A critical question, upon which there is no general agreement, is that of the time relation of the Ringold formation to the diastrophic activity that initiated the present erosion cycle.

[1] R. L. Lupher and Walter Warren, "The Asotin Stage of the Snake River Canyon near Lewiston, Idaho," *Jour. Geol.*, Vol. L (1942), pp. 866–81.

J. C. Merriam and J. P. Buwalda[2] believed that the Ringold was deposited in structural basins after the deformation. H. E. Culver[3] suggested that it was preceded by minor deformation and followed by major deformation, and Charles Warren[4] believes that deformation and deposition were largely contemporaneous. The writer believes that the Ringold, though early Pleistocene in age, is older than the major deformation because the sediments contain little erosional material from the basalts and extend into upland regions that have been trenched by the modern canyons. This conclusion leads to the thesis that the major deformation and dissection of the lava and, therefore, all lake and stream deposits of the present erosion cycle are of Pleistocene age.

The Ringold formation has not been certainly recognized in the vicinity of Lewiston Basin; a sedimentary deposit, largely eolian and commonly called the "Palouse soil" or "Palouse loess," lies upon the basalts in the Palouse Hills north of the basin, but little is known of its age or time range other than that the upper part is Pleistocene or Recent.

The major deformation of the lavas, presumably beginning early in the Pleistocene, produced the Columbia Basin, a major structural basin centering near the junction of the Snake and Columbia rivers and bordered by the Cascade, Blue, and Rocky mountains. In the Clarkston region (Fig. 1), well up on the eastern side of the basin, the lavas were warped and elevated so that streams of the present cycle carved many deep youthful canyons. Lewiston Basin at the junction of the Snake and Clearwater rivers is a minor structural and erosional depression about 18 miles long, 12 miles wide, and 2,000 feet deep (Fig. 2). The north side is a steep escarpment because the lavas, which are nearly horizontal beneath the plateau region to the north, bend down abruptly at angles of as much as 60°. The east-west axis of the syncline lies immediately south of Lewiston, close to the northern escarpment; the lavas rise southward at an average rate of about 3° to the Blue Mountains on the southwest and to Craig Mountain on the southeast. The structural closure of the asymmetric syncline is 1,700 feet, but the low central part has been eroded 300 feet deeper and somewhat widened by the Snake and Clearwater rivers, which debouch from deep canyons in the margins of the basins. Eight miles west of Lewiston the Snake leaves the basin through the northern escarpment, flowing at an altitude of 685 feet above the sea in a canyon 2,000 feet deep.

An early episode in the Pleistocene deformation and dissection of the lavas is recorded by the Asotin stage,[5] in which canyons across the growing Lewiston Basin syncline reached a maximum depth of 1,325 feet and then were nearly filled by lava flows, thus causing the streams to begin their work anew.

Most of the deformation and dissection came after the Asotin stage, and there is no known record of cessation of downcutting by streams until major canyons were nearly as deep as they are now; but before the end of the Pleistocene there came three episodes of proglacial deposition, each followed by dissection of the deposits and slight deep-

[2] "Age of Strata Referred to the Ellensburg Formation in the White Bluffs of the Columbia River," *Univ. Calif. Publ., Bull. Dept. Geol.*, Vol. X (1917), pp. 255–66.

[3] "Extensions of the Ringold Formation," *Northwest Science*, Vol. XI (1937), pp. 57–60.

[4] "Course of Columbia River in Southern Central Washington," *Amer. Jour. Sci.*, Vol. CCXXXIX (1941), pp. 221–22.

[5] Lupher and Warren, ftn. 1 (1942).

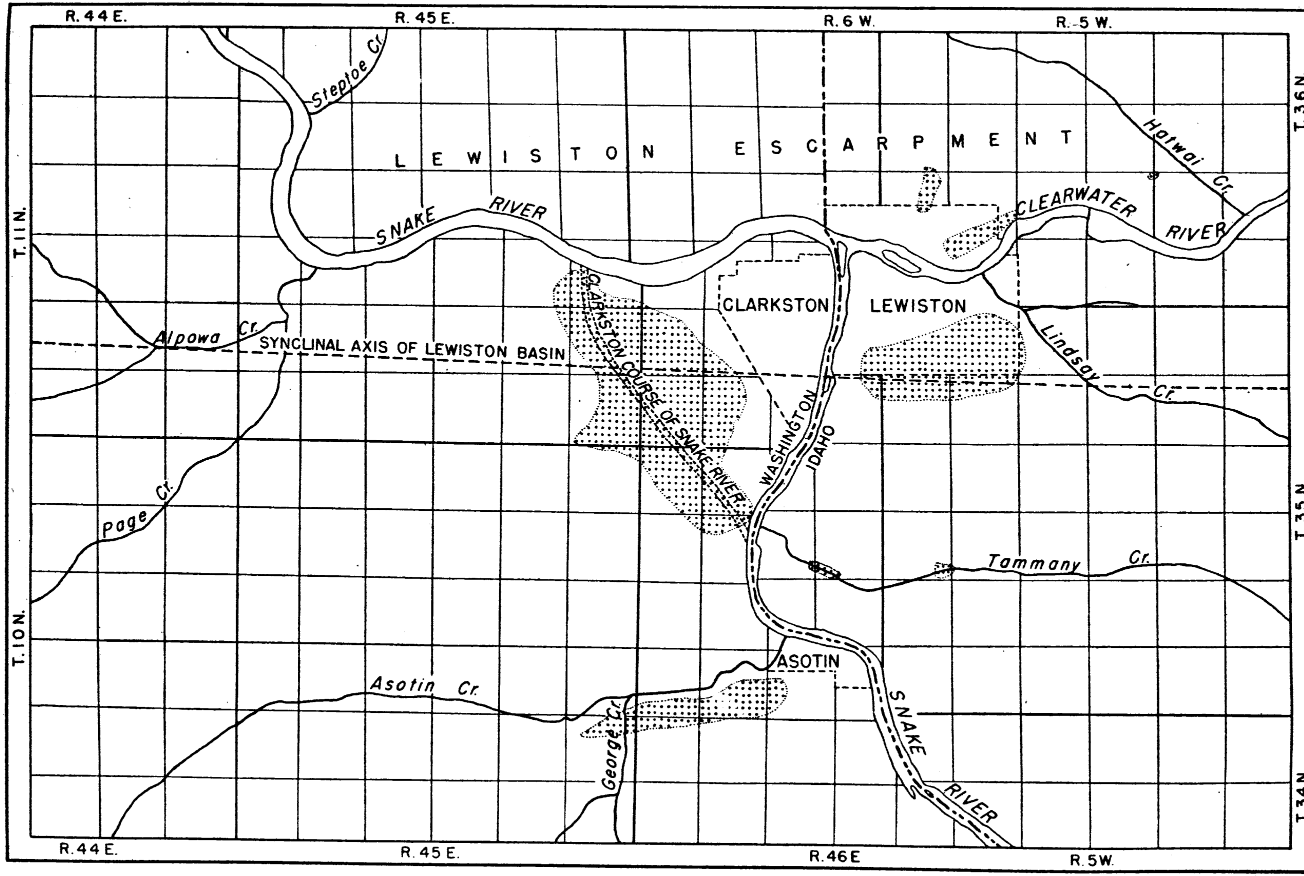


FIG. 1.—Map of the Lewiston Basin region. Approximate distribution of Clarkston fill remnants shown by stippling

ening of the original canyons. The first of these episodes is the Clarkston stage which is to be described in this paper; the second and third coincide with the much-discussed scabland history of the Wisconsin stage. The scabland history of central and southern Washington, as revealed by earlier writers, is too complex to be reviewed here; its principal effects in the Lewiston Basin region are found in local eroded tracts on the sides of canyons, discontinuous deposits of gravel and coarse sand,[6] and a more widespread mantle of silt and fine sand. The fine-grained deposits are the product of lake deposition; they are similar to, and probably contemporaneous with, the Touchet beds that were deposited in the Lake Lewis area of south-central Washington. Some of the coarse deposits were formed adjacent to ice jams which clogged the narrow lake in the manner advocated by Ira S. Allison;[7] others may record an earlier episode of normal aggradation.

[6] J Harlen Bretz, "Valley Deposits Immediately East of the Channeled Scablands of Washington," *Jour. Geol.*, Vol. XXXVII (1929), pp. 408–27, 505–9.

[7] "Flint's Fill Hypothesis of Scabland Origin," *Jour. Geol.*, Vol. XLIX (1941), pp. 54–73.

CLARKSTON STAGE

OUTLINE OF HISTORY

When the canyons were cut to within 50 feet of their present depths, a gravel fill began to accumulate along the Snake River and eventually reached a thickness of more than 400 feet. Some tributaries, because of steep gradients and adequate gravel loads, kept pace with the rising surface of the fill in the main canyon; but Tammany Creek was ponded, and there is some evidence that the Clearwater River was ponded in its lower course after aggrading to a depth of 320 feet.

FIG. 2.—View southward across Lewiston Basin. City of Clarkston on right; Clarkston gravel fill in former course of Snake River trends northwest beneath plowed field beyond; post-Clarkston course of Snake River in center; Clearwater River and Lewiston on left.

At the beginning of the Clarkston aggradation, Snake River followed a canyon northwest from a point opposite the mouth of Tammany Creek and was joined by the Clearwater 4 miles west of the present junction (Fig. 1). A ridge of resistant intracanyon lavas of the Asotin stage formed the interstream divide; and as the Snake River fill neared its maximum depth, the river periodically crossed the divide and poured gravels into the broad valley of the lower Clearwater in the structural low of the basin. The river was superposed on one of the

divide crossings during post-Clarkston dissection of the fill and cut a new canyon along its present course northward from Tammany Creek (Fig. 2).

The fill halted erosion in the deeper portion of the canyon, whereas normal lowering of slopes and stream gradients continued above the fill. Several rock terraces were cut at the highest level of the fill (Fig. 4). Subsequent erosion has removed most of the fill, and canyons now show a two-story effect: a rock-walled inner canyon below the original fill surface and more widely flaring upper slopes above the fill surface.

The cause of the Clarkston aggradation is unknown. However, it coincided, in some degree, with a glacial stage; and the common association of aggradation with proglacial conditions in other regions suggests that the Clarkston aggradation was induced by glaciation in the mountainous tracts drained by the Snake and Columbia rivers. The fill in Snake River canyon cannot be a valley-train deposit, in the sense of having been graded to a near-by glacier front. Presumably, the fill was caused either by increased load carried by tributary streams from widely distributed mountain glaciers in central Idaho and northeastern Oregon or, indirectly, by aggradation in the Columbia River valley at the mouth of the Snake. The diastrophic history of this part of the Pleistocene has not been elucidated; the possibility, therefore, of regional depression, or of faulting and warping west of Lewiston Basin, cannot be dismissed.

DESCRIPTION OF DEPOSITS

Stream deposits.—Most of the Clarkston fill is stream gravel in which are a few lenses of sand. The stream deposits are stained and cemented by iron oxide. Colors range mostly from light buff to brown, but some zones are so highly charged with oxide that they are orange or brick-red. Many pebbles are partly covered with a rough film of oxide. The degree of cementation is comparable to that of the Ringold formation of the earliest Pleistocene or to that of the Hood River conglomerate; stream gravels have stood with vertical and overhanging surfaces for many years without appreciable

FIG. 3.—Clarkston gravel in Clearwater valley, ½ mile south of Lewiston. Rough-surfaced pebbles near top are decomposed basalt.

modification, and some of the sand deposits are sufficiently indurated to allow shaping into fairly durable hand samples.

The stones in the fill brought by the Snake and Clearwater rivers are almost wholly in the pebble and cobble range (Fig. 3). Stones in the 2–6-inch range make up the bulk of the fills; but small pebbles are numerous and, at their lower limit, grade into a matrix of coarse- to medium-grained arkosic sand. A number of lenses of arkosic sandstone, a few

inches to a few feet in thickness, have been seen in the Clearwater valley fill. Most pieces of boulder size are rolled blocks from near-by canyonsides; a few are roundstones; and some are ice-rafted erratics.

It is difficult to determine the abundance of erratics because (1) loose blocks of foreign rocks lying on eroded Clarkston fill cannot be distinguished from erratics carried during the Lake Lewis episode of the Wisconsin stage, and (2) most of the angular blocks buried in Clarkston fill are basalt and are to be suspected of local derivation. However, two unmistakable Clarkston erratics were visible in April, 1944, in a gravel pit on South Eighteenth Street at the southeast margin of Lewiston. These two blocks are angular; one, of granodiorite, is 51 inches in diameter, and another, of biotite schist, is 24 inches in diameter. They were inbedded in well-sorted pebble and cobble gravel. A similar block of granodiorite was seen in this pit in 1941, but it was subsequently removed. The nearest possible source for these rocks is at least 25 miles east of Lewiston Basin and their presence in the fine gravels of the Clarkston fill can be explained only by ice-rafting. Though only three erratics are certainly identified, the number is comparable to that found in a like number of exposures in other proglacial stream deposits of southeastern Washington when a similar restriction—large angular blocks of foreign rock *in situ*—is imposed.

The gravels contain Tertiary basalts and pre-Tertiary igneous and metamorphic rocks in nearly equal amounts, though local basalt rubble and rolled blocks predominate near canyon walls. The pre-Tertiary components, having traveled considerable distances from their sources in central Idaho and northeastern Oregon, are mostly well rounded; their shapes range from nearly spherical to elongate and flattened (Fig. 3). The basalt components, having come from distant as well as local sources, range from well rounded to angular. The pre-Tertiary rocks are in large part the resistant residues of prolonged abrasion of waste from a varied metamorphic and igneous terrane. Though the streams tap large exposures of granodiorite, the pebbles that reached the Lewiston Basin region are largely medium- and fine-grained acidic rocks that are probably from aplite and complementary dikes, pegmatite and vein quartz, and resistant porphyries. From the metamorphic terrane have come mostly quartzites and hard flinty greenstones, the latter derived from basaltic and andesitic lavas, tuffs, agglomerates, and intrusives. Pebble counts[8] show that Snake River gravels, of Pleistocene, and Recent ages, are characterized by abundant greenstone pebbles, whereas the Clearwater River gravels are characterized by much light-colored quartzite and granodiorite; the difference is sufficient to produce more somber colors in the Snake River fill. The gravels have undergone marked weathering, so that many of the basalt and granodiorite roundstones crumble soon after exposure (Fig. 3).

The Clarkston gravels in minor canyons around Lewiston Basin are composed of local basalt stones in a matrix of basalt waste, clay, and a small amount of arkosic sand derived from interbasalt sedimentary layers. In the canyon of Asotin Creek, a vigorous tributary of the Snake, the materials range from boulder gravel to fine gravel, and many stones are well rounded; in the canyon

[8] W. O. McKenzie, "Pebble Counts in the Lewiston Basin" (unpublished thesis, State College of Washington, 1942).

of Alpowa Creek, a less vigorous tributary, the stones are smaller, less rounded, and mixed with rubble and finer unsorted basalt waste; along Lindsay Creek, a small intermittent stream, the fill is largely angular rubble, stones rounded by spheroidal weathering, and finer material—all largely the product of mass wastage from near-by hillsides.

The Clarkston gravels are most reliably distinguished from Wisconsin gravels and other younger deposits by greater consolidation, by iron oxide colors, and by abundant decomposed roundstones. In contrast, the younger gravels are gray or dark-gray, less consolidated, and relatively unweathered. Petrologic composition is not a reliable distinction because Clarkston deposits have been extensively re-worked.

Lake deposits.—In the bed of Tammany Creek, about a mile above its mouth, a fine, blue-gray lake clay is exposed to a depth of a few feet. Yellowish-gray, well-stratified siltstones and sandstones, 20 feet thick, lie upon the clay. I. C. Russell[9] noted these deposits and rightly explained them as lake deposits formed in a ponded tributary of the Snake below the level of Snake River fill. However, the prominent bar of gravel and silt that Russell noted in the mouth of the canyon is of post-Clarkston age, and the original Clarkston gravel dam is exposed beneath the younger deposits only in the upper end of a narrow Recent trench carved by Tammany Creek.

Well-stratified brown sandstone is exposed also on the north side of Lewiston Basin, 1 mile northeast of the Clearwater dam, at an altitude of 325 feet above the river. The sandstone may be the record of sluggish water where Clearwater River flowed upon a broad fill surface in the middle of the Lewiston Basin, or it may be the result of ponding of the river by the Snake River fill. Ponding of the Clearwater is suggested also by a rubble deposit along the highway grade to the plateau north of Lewiston. It lies at altitudes of 1,050–1,250 feet A.T., or 340–540 feet above Clearwater River. The rubble contains angular and slightly rounded pieces of basalt that range from small pebble size to blocks 16 inches in diameter. The matrix, which makes up about half of the deposit, is dark gray-brown clay. The maximum thickness exposed is about 20 feet. The deposit lies upon basalt and is overlain by Touchet beds. Somewhat similar rubble is interbedded with Touchet silt near the bottom of the grade, and the entire deposit may be a marginal facies of the Touchet lake deposits. However, the deposits at the higher level show a degree of cementation and staining that is comparable to that of the Clarkston deposits.

The highest known occurrence of Clearwater River gravels of Clarkston age, in a fill remnant south of Lewiston, is 1,020 feet A.T.; this is 170 feet below the high level of near-by Snake River fill. It is possible that Clearwater River, like Tammany Creek, was ponded by more rapid growth of Snake River fill and that the sand and rubble at high altitudes on the north side of the basin accumulated at the margin of a lake.

DISTRIBUTION AND ALTITUDES OF REMNANTS

Snake River canyon and tributaries.—The remnants of Clarkston fill in the Snake and Clearwater valleys are largely obscured by a mantle of younger lake and stream deposits. Therefore, the stippled areas on the map (Fig. 1) do not show the distribution of Clarkston

[9] "Geology and Water Resources of Nez Perce County, Idaho," *U.S. Geol. Surv. Water Supply Paper 53* (1901), p.73.

deposits in detail but only the general areas in which those deposits are reliably indicated by partial exposure and topographic expression.

The largest remnant of Clarkston gravels lies in the abandoned segment of the Snake River canyon southwest of Clarkston. The fill surface is intact for a distance of a mile in the middle portion and is dissected at each end. Touchet lake silt, which once covered the entire remnant to a depth of several feet, has been removed from the steep surfaces of dissected portions. The upper surface of the fill is somewhat concave, lying at 1,160 feet A.T. in the middle and rising to 1,190 feet A.T. at the margins. The bottom of the former canyon is not clearly exposed but apparently is not more than 30 feet above the present surface of Snake River and, therefore, between 40 and 50 feet above the present bed of the river. These figures give a maximum vertical range of 450 feet from the highest marginal portions of the fill to the buried canyon bottom.

Most of the fill has been removed from the deep canyon below Lewiston Basin, though a few small remnants were noted in tributary gulches for a distance of 20 miles northwest of the basin. Remnants between Steptoe and Nisqually John creeks have flat tops that lie at 1,030 feet A.T., or 390 feet above present river-level. If the flats correspond to the highest level of the fill, a downstream slope of 12.8 feet per mile is indicated for the fill surface between Clarkston and Nisqually John Creek. The present gradient of the river, over the same reach, is 3.3 feet per mile.

Large remnants of the fill are present along lower Asotin Creek on the south side of Lewiston Basin and are well exposed in cuts along roads to the uplands south of the canyon. The upper surface of the fill is preserved by a veneer of stream gravel upon a prominent rock terrace (Fig. 4). The surface lies at 1,240 feet A.T. near the mouth of the canyon and rises to 1,335 A.T. 2.5 miles upstream, thus indicating a downstream slope of 38 feet per mile. The Clarkston lake sediments on lower Tammany Creek rise to 950 feet A.T. near the mouth of the creek and to 1,150 feet A.T. 3 miles upstream on a terrace that marks the former lake margin.

FIG. 4.—View westward up lower Asotin Creek, showing partly dissected rock terrace surfaced with Clarkston gravel. Note inner canyon below fill surface and subdued profiles above.

Clearwater valley and tributaries.—Prior to the Clarkston stage, Clearwater River flowed westward near the structural low of Lewiston Basin and joined the Snake 3.5 miles west of the present junction. The river had enlarged and deepened the downwarp, so that a structural and erosional depression, more than 3 miles wide and 6 miles long, lay below the upper limit of the near-by Snake River fill. There is no proof at present that this depression was filled with stream gravel to the high level of the Snake River fill.

The somewhat unusual nature of the elevated masses of Pleistocene sediments upon which the twin cities of Lewiston and Clarkston are built (Fig. 2) was noted and described by Bretz:

> The town and orchard tracts of Clarkston, Washington, are located on a sloping terrace-like fill composed of sand and gravel with prominent deltaic foresets. The deposit is a broad semimound resting against the southwest wall of the valley, its surface descending from the highest part roughly 50 feet per mile, down the valley, across the valley, and up the valley. Its upstream margin has been cliffed by undercutting of the Snake. This semimound is an impossible shape for a terrace top or a delta top, and the slopes are far too steep for either feature in a valley as broad as the Snake. There is no tributary drainage entering here from the higher land to the southwest, to which it might be referred as an alluvial-fan deposit.
>
> Pits in several places afford some idea of the nature of the material. Two different deposits underlie the tract: an older stained and weathered gravel beneath an unweathered younger deposit in which Columbia basalt constitutes about 80 per cent.[10]

The deposit beneath Lewiston has a closer resemblance to a terrace, especially when viewed from the west or north, where it has been undercut by the Snake and Clearwater rivers; but the top surface is noticeably rounded, with the high central point rising to an altitude of 850 feet A.T., or 140 feet above the adjacent river (Fig. 2). The top and undercut margins reveal only post-Clarkston deposits, but Clarkston gravels appear from beneath the younger deposits on the south side of the mound. South and southeast of Lewiston, a large dissected mass of gravel fill, partly obscured by a mantle of younger deposits, continues up the side of the basin to an altitude of 1,020 feet A.T. It seems probable that the Lewiston mound and the Clarkston semimound are largely erosional remnants of Clarkston gravel which are now largely hidden by younger surficial deposits.

The surficial deposits include partly cemented gray gravels intermediate in age between the Clarkston and the latest Pleistocene Touchet beds and scabland deposits. The intermediate gravels are well exposed along the north and northwest margin of the Lewiston mound. They resemble Clarkston deposits only in an appreciable degree of cementation, and they lack the iron oxide stain and weathered stones that characterize the Clarkston. Moreover, the gravels are of Snake River origin and were deposited near the bottom of the Clearwater valley after much of the original Clarkston fill had been removed. Similar gray gravels, 5–10 feet thick, lie between Clarkston remnants and the undulating mantle of Touchet lake silt at more than 1,000 feet A.T. on the south and southeast sides of Lewiston. These intermediate gravels are largely re-worked Clarkston gravels, probably left upon valley and canyon slopes during the dissection of Clarkston fill. Similar conditions may be repeated near Alpowa Creek and at several places in the canyon below Lewiston Basin, where alluvial deposits rise 50–135 feet above the

[10] Bretz, pp. 419–20 of ftn. 6 (1920).

river. Between Steptoe and Nisqually John creeks and at Davis Bar, near Almota, they form distinct terraces about 110 feet above the river; other deposits are of barlike or semimound aspect. The larger masses reach about the same altitude above the river as the top of the Lewiston mound, thus suggesting a considerably eroded post-Clarkston terrace. Exposures are shallow and reveal only gray, partly cemented gravel and later Pleistocene deposits; so it is not known whether the remnants represent an aggradational episode in the Clarkston-Touchet interval or a surface cut upon Clarkston fill and mantled by re-worked materials.

TOPOGRAPHIC EFFECTS ON COLUMBIA RIVER BASALTS

Two-story canyons.—The most pronounced effect of the Clarkston stage upon the regional topography was the development of two-story canyons. These canyons are remarkably similar to those formed by rejuvenation, but in this case they are related to aggradation. The alluvial fill of the Snake River and most of its tributaries eventually buried the lower portions of the youthful canyons to depths of more than 400 feet. The canyonsides probably were undercut by the streams during the upbuilding of the fill and perhaps also during its dissection; but the effect of the fill was primarily to bury and, in a sense, to fossilize the steep and rocky lower portions of the canyons for a long time. In the Clearwater canyon and its tributaries, alluvial fill, perhaps in combination with impounded water, produced the same result.

The fill must have remained at or near its maximum depth for a long time, because the canyon slopes above were considerably lowered by erosion and eventually blanketed by soil and vegetation. Likewise, the gradients of steams tributary to the Snake and Clearwater were progressively lowered upstream from the fill, which, as a rule, did not extend back more than 10 miles from the main river. Post-Clarkston dissection of the fill excavated the buried portions of the canyons and so revealed their steep rocky slopes in sharp contrast with the gentler, partly soil-covered slopes above.

The two-story character is prominent in the Snake River canyon northwest of Lewiston Basin and can be readily seen on the United States Geological Survey topographic map of the Pullman, Washington, quadrangle. The inner gorge does not owe its origin entirely to burial and exhumation, for it was locally eroded in its lower portion during the Wisconsin stage; and undercutting on the outsides of river bends, still in progress, has developed local precipitous slopes that rise even higher than the gorge. However, between the undercut slopes, even on the insides of river bends, are numerous steep slopes that rise from 350 to 400 feet above the present river and coincide at their upper limit with the highest level of Clarkston fill remnants. The canyonsides above the inner gorge and undercut slopes are noticeably less steep, and their contrast with the rocky cliffs of the inner gorge is heightened by a partial cover of soil and vegetation. A series of profiles, constructed from the topographic map of the Pullman quadrangle, shows an average slope of 32° on the inner gorge and 14° on the outer canyon.

In Asotin Creek canyon, south of Clarkston, a narrow inner gorge is 475 feet deep at the mouth of the canyon but decreases in depth upstream; and about 7 miles from the Snake River the

gorge ends, and the entire canyon has widely flaring soil-covered slopes. The stream gradient lowers abruptly as the canyon widens, and the knickpoint probably marks the upstream terminus of the fill. Similar conditions are encountered on Alpowa Creek. East of the basin the two-story character is well shown by the canyons of Clearwater River and Potlatch Creek; Cottonwood Creek canyon repeats the Asotin Creek–Alpowa Creek features.

Proof that the two-story canyons were caused by the Clarkston aggradation cannot rest entirely upon the coincidence of fill and inner canyon, for there is the possibility of a fortuitous coincidence of Clarkston fill surface with the knickpoints of a pre-Clarkston two-cycle valley formed by rejuvenation after an earlier valley stage. The best evidence for the aggradation thesis is the failure of inner gorges to continue upstream beyond the fill that lay in the lower portions of large canyons tributary to the Snake and Clearwater.

Rock terraces.—The "two-cycle" appearance of the canyon profiles is enhanced by rock terraces at the upper limit of the inner gorges. The terraces resemble the strath remnants of an early valley that are commonly preserved long after rejuvenation of a stream, but they are rock terraces cut by streams running at the upper level of Clarkston fill. The best example seen (Fig. 4) lies at the mouth of Asotin Creek canyon and extends upstream for 3 miles; it corresponds in altitude to the upper level of Clarkston fill and is partly covered by gravel. A similar but smaller terrace lies immediately west of Wawawai Creek in the Snake River canyon. Elsewhere along the Snake and Clearwater rivers the terraces are not large, but many spurs that come down the widely flaring outer canyon are beveled by horizontal or gently inclined surfaces at their junction with the inner gorge.

AGE

That the Clarkston stage coincided, to some degree, with a glacal stage of the Pleistocene is assured by large ice-rafted erratics in the stream deposits. It seems certain that it occurred long after the beginning of the Pleistocene but prior to the Wisconsin stage. It is not yet possible to estimate the position of the stage with reference to the Mississippi Valley sequence because of uncertainties regarding both number and ages of Northwest glacial stages; for the present it must be sufficient to locate the stage only with reference to local history.

The belief that the Clarkston stage occurred well within the Pleistocene is based upon the evidence that the major deformation, which initiated the present erosion cycle, came after the deposition of the early Pleistocene Ringold formation. Therefore, considerable time must be allotted for the Ringold deposition, the deformation and associated canyon-cutting, and, in the Lewiston Basin region, a repetition of canyon-cutting after lavas had nearly filled canyons of the Asotin stage. It is, perhaps, a late origin of the modern topography that has prevented recognition of early Pleistocene glacial stages in the Columbia Basin region.

An upper time limit for the Clarkston stage is set by the Touchet beds and little-consolidated scabland gravels and sands which mantle erosional remnants of Clarkston fill. It is generally agreed[11]

[11] Bretz, "The Age of the Spokane Glaciation," *Amer. Jour. Sci.*, 5th ser., Vol. VIII (1924), pp. 336–42; Richard F. Flint, "Origin of the Cheney-Palouse Scabland Tract, Washington," *Bull. Geol. Soc. Amer.*, Vol. XLIX (1938), p. 468; Allison, p. 71 of ftn. 7 (1941).

that these younger deposits are of Wisconsin age. Allison[12] recognizes a twofold division of Wisconsin deposits near the mouth of Snake River; the older unit is represented by the high gravel terrace corresponding to the gravel dam[13] at the mouth of Palouse River, and the younger by Touchet beds and coarse scabland deposits of the Lake Lewis–ice-jam episode. He believes[14] that both units are younger than Clarkston deposits which are present in the same general region.

That the Clarkston deposits are older than Wisconsin is indicated also by their advanced degree of cementation, weathering, and iron oxide staining. Cementation of Clarkston deposits is comparable to that of early Pleistocene deposits, such as the Ringold formation, and many of the stream-worn cobbles and boulders are rotten to the core. In degree of weathering they resemble the Peshastin till described by Ben M. Page[15] as the first record of three successive Pleistocene glaciations in Wenatchee Valley, Washington. In contrast, the Wisconsin deposits seen by the writer and those described by Bretz,[16] Flint,[17] and Allison[18] are little consolidated, essentially unweathered, and lack a prominent oxide stain. Other evidences of antiquity are found in the length of time that would be required for the growth of the fill and the marked lowering of the outer canyon slopes while the fill remained in the canyon bottoms. Furthermore, the erosional interval between the Clarkston stage and the beginning of the Wisconsin deposition was of considerable magnitude. The fill, over 400 feet deep, was largely removed from Lewiston Basin and neighboring canyons; the Snake River cut a new course across the basalt ridge south of Lewiston and elsewhere cut its channel about 30 feet deeper into the basalt bedrock. Post-Wisconsin erosion is insignificant by comparison, for the thin mantle of Touchet lake silt is intact over much of the Lewiston Basin.

The conclusion indicated is that the Clarkston episode is a distinct proglacial stage in the middle part of the Pleistocene. The implication, of course, is that it is equivalent to either the Kansan or the Illinoian stage of the Mississippi Valley sequence, but judgment on that question must be deferred until further studies are made.

Acknowledgment.—The writer is indebted to Dr. Ira S. Allison, who read the manuscript of this paper and gave many helpful comments.

12 Pp. 59–71 of ftn. 7 (1941).

13 Flint, pp. 507–10 of ftn. 12 (1938).

14 Personal communication (1944).

15 "Multiple Alpine Glaciation in the Leavenworth Area, Washington," *Jour. Geol.*, Vol. XLVII (1939), pp. 785–815.

16 Ftn. 6 (1929).

17 Ftn. 11 (1938).

18 Ftn. 7 (1941).

A MAJOR BURIED VALLEY IN EAST-CENTRAL ILLINOIS AND ITS REGIONAL RELATIONSHIPS[1]

LELAND HORBERG
Illinois State Geological Survey, Urbana

ABSTRACT

A large buried valley, tributary to the well-known bedrock valley along the Illinois River, has been traced eastward across central Illinois to the Indiana state line, and continuation into Indiana and beyond is indicated by well records and outcrop data. Detailed studies are confined largely to Illinois, and the valley is herein named "Mahomet Valley" after a locality in Champaign County, Illinois. Mahomet Valley is considered preglacial, as Kansan, Aftonian, and possibly Nebraskan deposits occur within the channel. After the advance of the Kansan glacier the valley probably ceased to function as a major drainage line; and by the end of Illinoian time the valley was so completely filled with drift that the Sangamon interglacial plain continued across it without interruption.

A new working hypothesis favored by the writer is proposed, namely, that Mahomet Valley represents the lower course of Teays River, a preglacial master-stream which probably had its source near the eastern scarp of the Blue Ridge in North Carolina; flowed westward across Ohio, northern Indiana, and central Illinois; and finally discharged into the Gulf Embayment through bedrock valleys now generally occupied by the present Illinois and Mississippi rivers.

INTRODUCTION

The geological studies made by the Worthen Survey of Illinois revealed that at several points in central Illinois the bedrock occurred at elevations much lower than at adjacent localities. The distribution of these low points led F. H. Bradley[2] to postulate that a preglacial valley extended southward from Lake Michigan through Kankakee and eastern Iroquois counties into Champaign County and thence northwestward under the city of Bloomington into the Illinois Valley in southern Tazewell County. It is now known that these low points lie within independent preglacial drainage systems.

Frank Leverett[3] confirmed the presence of low bedrock elevations in east-central Illinois and suggested possible relations to the preglacial courses of the Kaskaskia, Wabash, and Illinois rivers. In a regional summary in 1910, H. M. Clem[4] suggested the presence of a "spur" connecting Illinois and Wabash bedrock valleys; and in 1931, T. E. Savage[5] definitely related the preglacial drainage of the region to the Illinois bedrock valley. Within recent years L. E. Workman and George E. Ekblaw, of the Illinois State Geological Survey, in unpublished maps and cross sections outlined the eastern margin of the valley in Champaign County and made the first subsurface interpretation of the glacial deposits within the area.

The name "Mahomet" is herein proposed for the major bedrock valley crossing the area because near the village of Mahomet in western Champaign County

[1] Published with the permission of the chief of the Illinois Geological Survey, Urbana, Illinois

[2] "Geology of Kankakee and Iroquis Counties," in *Geology and Paleontology* ("Ill. Geol. Surv.," Vol. IV [1870]), pp. 226–40; "Geology of Champaign, Edgar and Ford Counties," *ibid.*, pp. 266–75.

[3] "The Preglacial Valleys of the Mississippi and Its Tributaries," *Jour. Geol.*, Vol. III (1895), pp. 744–57; "The Illinois Glacial Lobe," *U.S. Geol. Surv. Mono.*, Vol. XXXVIII (1899), pp. 654–64, 701–7.

[4] "The Preglacial Valleys of the Upper Mississippi and Its Eastern Tributaries," *Proc. Ind. Acad. Sci., 1910* (1911), pp. 335–52.

[5] "On the Geology of Champaign County," *Trans. Ill. Acad. Sci.*, Vol. XXIII (1931), pp. 444–45.

three wells penetrate bedrock at low elevations and determine the position of the deep part of the channel.

The present study is an outgrowth of a ground-water study of Pleistocene aquifers in central Illinois (Fig. 1), in which all available well records were examined, detailed studies were made of about seventy-five sets of well cuttings, and a contour map of the bedrock surface (Fig. 2) was compiled. Data for the bedrock-surface map of western Indiana were compiled from the literature and from well records provided by Wallace W. Hagan, formerly of the Indiana Division of Geology.

The thickness of the glacial drift in the area ranges from 0, where bedrock is exposed at the surface, to over 450 feet, where moraines of Wisconsin age cross deep bedrock valleys. Few water wells reach bedrock where the drift is over 200 feet thick,[6] whereas numerous wells reach it where the drift is thin. For this reason considerable detail of relief is usually shown on the bedrock uplands, but only the general outlines of the major valleys are revealed.

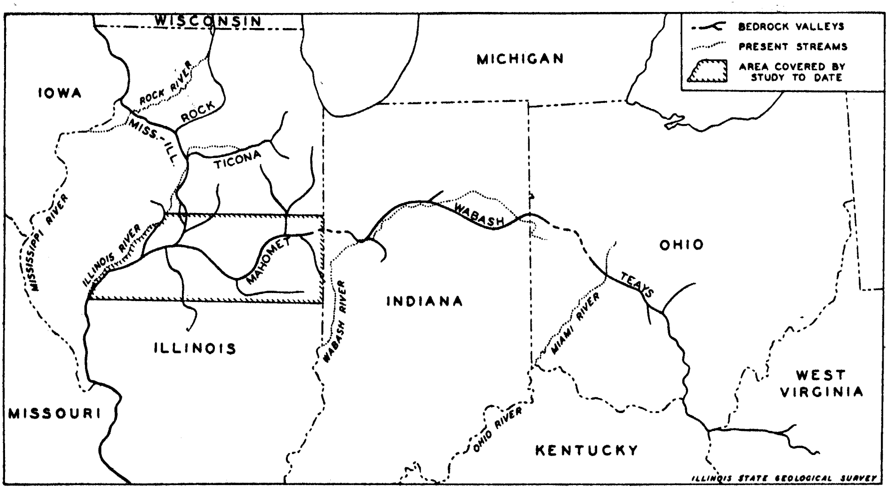


FIG. 1.—Map showing area studied in Illinois and proposed Teays drainage system. (Adapted in part after Fidlar, Tight, and Ver Steeg.)

DESCRIPTION

According to the present study, Mahomet Valley enters the state near the southeastern corner of Iroquois County and with a broad southward loop continues westward for 120 miles to enter the Illinois bedrock valley in southern Tazewell County (Fig. 2). Bedrock elevations along the valley are less than 400 feet above sea-level, or 200–300 feet lower than elevations on the adjoining bedrock uplands. The average depth of the valley appears to be about 200 feet, and in general the valley lies between the 300- and the 500-foot contours. The width of the inner portion of the valley lying below elevations of 450 feet is about 4 miles near the east state line, 5 miles in central Piatt County, and about 15 miles in DeWitt County. Although the valley-

[6] Mr. C. F. Stiegman, a water-well driller at Paxton, states that he has drilled wells along Mahomet Valley for twenty-five years without encountering bedrock in any of them.

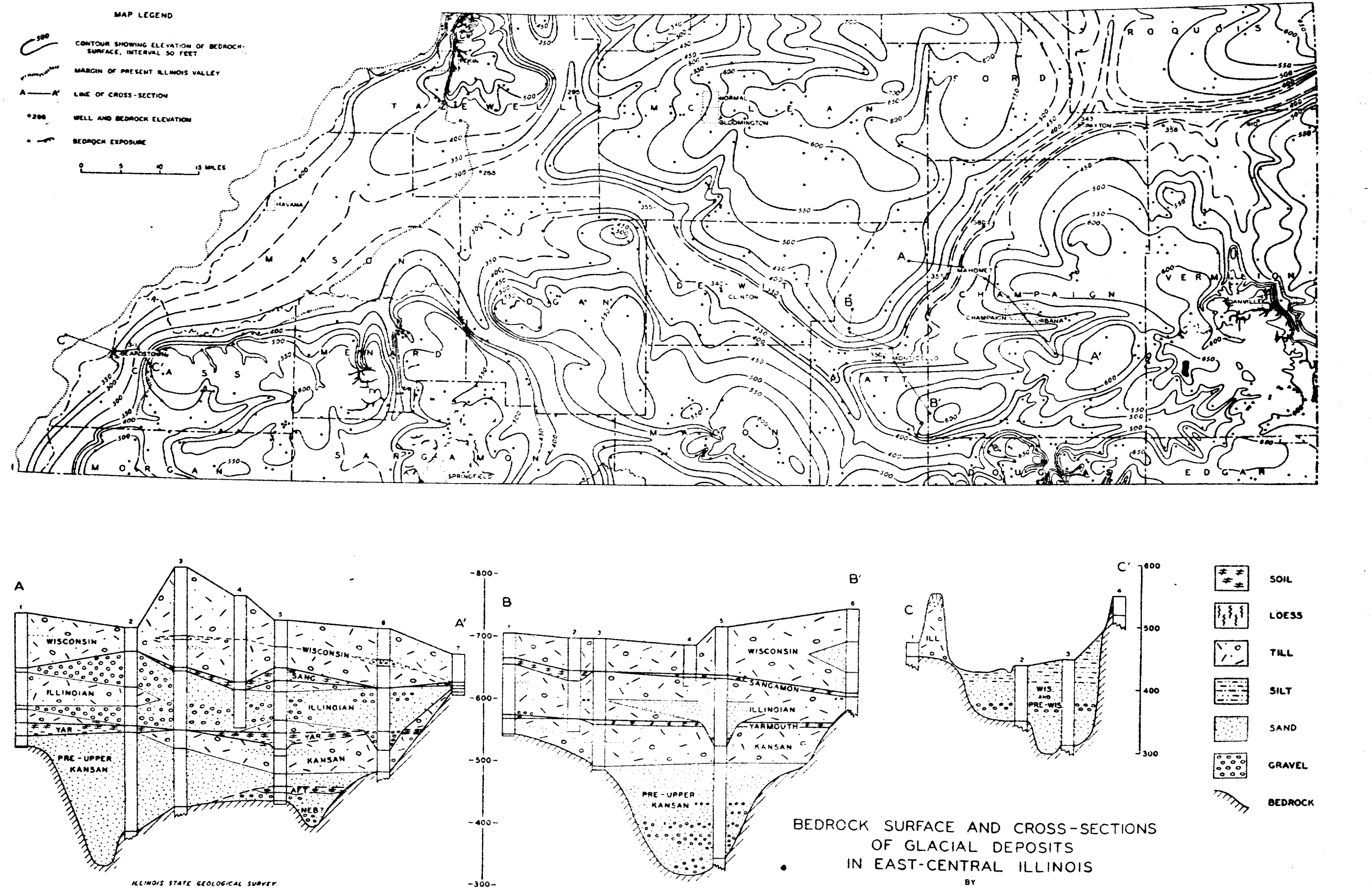

MAP LEGEND
CONTOUR SHOWING ELEVATION OF BEDROCK-SURFACE, INTERVAL 50 FEET
MARGIN OF PRESENT ILLINOIS VALLEY
LINE OF CROSS-SECTION
WELL AND BEDROCK ELEVATION
BEDROCK EXPOSURE
0 5 10 15 MILES
TAZEWELL
MCLEAN
FORD
IROQUOIS
MASON
LOGAN
DEWITT
CHAMPAIGN
VERMILION
CASS
MENARD
SANGAMON
MACON
PIATT
DOUGLAS
EDGAR
MORGAN
PEKIN
HAVANA
NORMAL
BLOOMINGTON
CLINTON
MAHOMET
CHAMPAIGN
URBANA
DANVILLE
MONTICELLO
SPRINGFIELD
BEARDSTOWN
PAXTON
WISCONSIN
ILLINOIAN
YAR
PRE-UPPER KANSAN
SANG
KANSAN
AFT
NEB?
SANGAMON
YARMOUTH
ILL
WIS. AND PRE-WIS.
SOIL
LOESS
TILL
SILT
SAND
GRAVEL
BEDROCK
BEDROCK SURFACE AND CROSS-SECTIONS
OF GLACIAL DEPOSITS
IN EAST-CENTRAL ILLINOIS
BY
LELAND HORBERG, MAY 1944
ILLINOIS STATE GEOLOGICAL SURVEY

FIG. 2

walls cannot be drawn sharply in most places because of the lack of detailed data, a notable widening is indicated for the downstream portion of the valley.

The descent of the valley floor appears to be gradual; and, when estimated on the basis of a minimum elevation of 280 feet above sea-level along Illinois bedrock valley and the elevation of 300 feet above sea-level at Oxford, Indiana,[7] an average gradient of 1.65 inches per mile is obtained. This is a descent of 20 feet in 145 miles.

A mature stage of development of the valley is indicated by its relative width and depth and by the wide distribution of low elevations in DeWitt County, which suggest the presence of a flood plain. There is also a suggestion that the valley may have been eroded during two cycles so that the inner valley is entrenched below a broad outer valley. This is evidenced by the pronounced break in slope below the 550- and the 500-foot contours and by the absence of comparable low elevations outside the inner valley.

A single major tributary from the north enters Mahomet Valley near Paxton. This valley appears to have its source along the margin of the Niagaran escarpment in northeastern Illinois. Important tributaries from the south enter Mahomet Valley north of Danville, west of Monticello, and in western Logan and northern Menard counties, which is in opposition to the general slope of the present drift plain.

Bedrock uplands bordering the valley range in elevation from 720 to 550 feet above sea-level, with the most extensive areas falling between the 550- and 600-foot contours. These uplands are parts of preglacial watersheds that separated River Mahomet from River Ticona[8] to the north, from upper Mississippi drainage to the northwest, and from Wabash and lower Mississippi drainage to the south (Fig. 1).

Maximum total relief for the area is about 460 feet, with the lowest elevations, 280–290 feet above sea-level, along Illinois bedrock valley and the highest elevations, about 720 feet above sea-level, on the bedrock upland in northeastern McLean County.

In the absence of closely spaced data along the upper course of the valley two alternative interpretations may be considered: (1) a low divide near the northern boundary of Champaign County may have separated Mahomet Valley from another valley east of Paxton, which drained eastward rather than westward; (2) there may be a divide near the state line so that Mahomet Valley did not extend into Indiana. By both of these interpretations major valleys end abruptly without important headwater tributaries, thus failing to satisfy physiographic requirements. The first alternative is further discounted by the northwest trend of the tributary valley north of Danville, indicating drainage to the west, and by a record showing bedrock less than 380 feet above sea-level in the northwest part of Champaign County. The major objection to the second alternative is the low bedrock elevations in southern Benton County, Indiana (Fig. 4). In view of these facts the writer's interpretation, shown in Figure 2, will be assumed in subs quent descriptions.

RELATION TO PRESENT TOPOGRAPHY

The present topography of the area is controlled almost entirely by moraines of the Wisconsin glacial stage, which bear no direct relation to the bedrock

[7] Leverett, p. 757 of ftn. 3 (1895).

[8] H. B. Willman, "Preglacial River Ticona," *Trans. Ill. Acad. Sci.*, Vol. XXXIII (1940), pp. 172–75.

topography and cross Mahomet Valley at various angles without change in trend or elevation. These moraines belong to the Tazewell substage, and in succession northeastward from the outermost are the Shelbyville, Cerro Gordo, Champaign, Bloomington, Normal, and Chatsworth.

The western part of the area lies west of the Wisconsin drift margin (Fig. 3) and includes uplands underlain by loess-covered Illinoian drift and a broad bottom land along the present Illinois River (Fig. 2). The lowland, which is a striking feature of the middle Illinois Valley, coincides with an extensive bedrock lowland developed at the confluence of the Mahomet and Illinois bedrock valleys and four important tributary bedrock valleys.

RELATION TO BEDROCK

Mahomet Valley cuts across regional structural trends and, from east to west, crosses the western Indiana syncline, the LaSalle uplift, and the northern part of the Illinois basin. The rocks underlying this area are largely nonresistant Pennsylvanian shales, although limestone and sandstone beds locally form thin units of greater resistance to erosion.

The major feature due to differential erosion is the bedrock lowland along the Illinois River and the related narrows at Beardstown (cross section *C–C′*, Fig. 2). The narrows resulted from entrenchment in more resistant Mississippian limestones, which are exposed along the river at this point; and the lowland may be attributed to lateral planation of weaker Pottsville and Carbondale strata upstream from this local base-level. Other features of the preglacial surface which locally appear to reflect bedrock lithology are: (1) the broad ridge in western Ford County may be due to Devonian-Silurian limestones along the LaSalle uplift; (2) the small upland in southwestern Vermilion County represents an outlier of LaSalle limestone; (3) the narrow ridge in north-central Douglas County is a reflection of Devonian-Silurian limestones along the crest of the LaSalle uplift; (4) the valley-wall in northern Menard and northwestern Logan counties may be partly the result of control by Pennsylvanian limestones above No. 6 coal.

DESCRIPTION OF THE VALLEY-FILL

Eleven units of Pleistocene deposits have been identified in the area:

11. Post-Shelbyville (Wisconsin) till and outwash
10. Shelbyville (Wisconsin) till
9. Sangamon soil and alluvium
8. Upper Illinoian till
7. Middle Illinoian sand and gravel
6. Lower Illinoian till
5. Yarmouth soil and alluvium
4. Kansan till
3. Kansan sand and gravel
2. Aftonian alluvium
1. Nebraskan (?) sand and gravel

Within this sequence significant unconformities occur at the base of the Kansan sand and at the bases of the Yarmouth and the Sangamon interglacial deposits. These unconformities are responsible for major variations in the succession below the Sangamon soil zone; the Wisconsin tills form a relatively regular unit, in which Shelbyville and post-Shelbyville divisions are usually recognizable. An outstanding feature of the pre-Wisconsin deposits is the dominance of water-laid silts, sands, and gravels within Mahomet Valley in contrast to glacial till, which is the dominant material along the margins of the valley and under the adjoining uplands (cross sections, *A–A′* and *B–B′*, Fig. 2).

RELATIONS OF THE SANGAMON AND THE YARMOUTH SOIL ZONES

The Sangamon plain below Wisconsin drift has been reconstructed for a part of the area on the basis of about 200 well records in which buried soil was either logged by the driller or was determined from sample cuttings (Fig. 3). Considering the minor errors possible in logging and in determining the location and elevation of wells, the close agreement of data-points for any given part of the map is noteworthy. The plain slopes gently away from the bedrock upland in McLean County and crosses Mahomet Valley without significant change in gradient. In Macon, Piatt, and western Champaign counties most of the elevations fall between 620 and 640 feet above sea-level. The surface has an average gradient of about 5 feet per mile and varies in elevation from 760 to 590 feet above sea-level. In northwest Champaign County a shallow sag in the plain lies approximately over Mahomet Valley, but elsewhere there is no coincidence with the bedrock surface. Upon this plain were spread the Wisconsin drift sheets and their moraines.

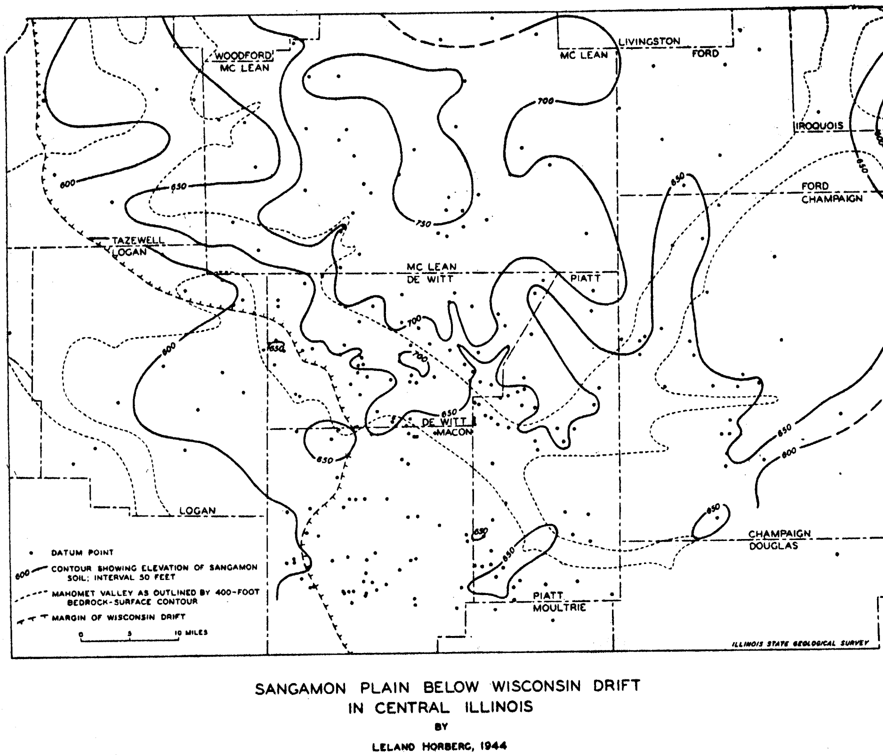


FIG. 3

The Yarmouth surface is not so well known as the Sangamon, but it appears to cross Mahomet Valley at fairly uniform elevations (cross sections A–A' and B–B', Fig. 2). About 60 wells in the area covered by Figure 3 encounter the horizon and indicate that the undissected

plain was generally parallel with the Sangamon surface. Highest elevations occur over the bedrock upland in McLean County, and from that area the Yarmouth surface slopes outward in all directions. Elevations range from 670 to 514 feet above sea-level, most of them falling between 550 and 600 feet.

INTERPRETATION OF LOCAL DRAINAGE HISTORY

The oldest feature of the bedrock surface is represented by the upland surface, which crosses the structures of the area and slopes southwestward from elevations of about 600–500 feet above sea-level. The most extensive parts of this surface appear to lie between the 550- and the 600-foot contours. In the northern part of the area in McLean and eastern Iroquois counties higher portions of the upland rise to a uniform level of about 650 feet and have restricted summits at elevations over 700 feet. In northwestern Illinois a summit erosion surface, called the "Dodgeville peneplain,"[9] has been recognized. This surface slopes southward from an elevation of about 900 feet in the Driftless Area to an elevation of about 600 feet in the Starved Rock region in LaSalle County.[10] Bedrock valleys, eroded 200–300 feet below this upland, are pre-Kansan and probably preglacial in age. The elevation, slope, and dissection of the bedrock uplands in east-central Illinois suggest their correlation with the Dodgeville surface.

[9] A. C. Trowbridge, "The Erosional History of the Driftless Area," *Iowa Univ. Studies*, 1st ser., No. 40 ("Studies in Nat. Hist.," Vol. IX, No. 3 [1921]), pp. 1–127; R. E. Bates, "Geomorphic History of the Kickapoo Region, Wisconsin," *Bull. Geol. Soc. Amer.*, Vol. L (1939), pp. 819–80.

[10] H. B. Willman and J. N. Payne, "Geology and Mineral Resources of the Marseilles, Ottawa, and Streator Quadrangles," *Ill. Geol. Surv., Bull. 66* (1942), pp. 204–5.

Mahomet Valley and its tributaries were eroded below the upland surface in pre-Aftonian time, as Aftonian and possibly Nebraskan deposits have been identified within the valley in cuttings from wells at Urbana, Champaign County (Fig. 2, cross section *A–A′*, well No. 5), and in southwestern McLean County. In both localities three soils are recognizable, the lowermost or Aftonian being underlain by sand and gravel. The age of the basal sand and gravel is uncertain, and it is considered Nebraskan rather than Aftonian largely because of the absence of humus, the pronounced break at the top of the deposit, and its general similarity to known glacial, rather than interglacial, deposits. Valley cutting thus appears to have been completed by preglacial Pleistocene time. Later modifications of the bedrock surface in the area were probably brought about largely by drainage diversions and only to a minor degree by true glacial corrasion. Glacial erosion by Wisconsin ice was certainly negligible, as there are few instances where the surface drift is not underlain by older glacial deposits.

The dominant glaciofluvial character of the pre-Wisconsin deposits within Mahomet Valley indicates that the valley remained an active drainage line until late Illinoian time. During the early Pleistocene the channel was probably open and cleared of fill, as the pre-upper Kansan sand and gravel in several places rests directly on bedrock (cross sections *A–A′* and *B–B′*, Fig. 2). However, after the deposition of this material, the valley was progressively filled with glacial and interglacial deposits so that by Sangamon time (possibly even by Yarmouth time) it had ceased to function as an important drainage-way and the Sangamon plain crossed it without interruption.

The available evidence indicates that the initial drainage diversion leading to the abandonment of the valley was caused by the advance of the Kansan glacier and that the valley continued only as a minor channel-way during Yarmouth and Illinoian time. The possibility of diversion in pre-Kansan time is opposed by the occurrence of Aftonian and older deposits at elevations between 450 and 500 feet above sea-level within the valley and by the stratified character of all the deposits within the valley that lie below the upper Kansan till. The alternative of diversion by the Illinoian glacier finds some support in the widespread occurrence of middle Illinoian sand along the valley. However, the base of this sand has an elevation of about 550 feet above sea-level, which is 250 feet above the valley-floor and close to the level of much of the upland. This relation, together with the uniform elevation of the Yarmouth soil, suggests that the valley in Illinoian time was a broad sag which followed the general course of Mahomet Valley and received Illinoian outwash but was not an important through-valley. This view is further attested by the fact that the Yarmouth and Sangamon deposits consist largely of peaty soil and alluvial silt and fine sand, most of which probably represents wash from adjacent gentle till slopes.

With the advance of the Wisconsin glacier across the Sangamon plain, all vestiges of the old valley were obliterated, and there is nothing in the present landscape to suggest its existence.

RELATION TO REGIONAL PRE-GLACIAL DRAINAGE

Numerous well records in southern Benton County, Indiana, suggest that Mahomet Valley continues eastward to join the bedrock valley along the present Wabash River near LaFayette. This interpretation (Fig. 4) is based on well records and bedrock-exposure data compiled largely from the literature and is thus subject to important revisions, although the amount of published data and their agreement are notable.

Low bedrock elevations in southern Benton County were first noted by S. S. Gorby[11] in 1866 and were subsequently verified by Leverett.[12] Three interpretations of these low bedrock elevations have been proposed: (1) the preglacial valley at LaFayette continues west past Oxford (Fig. 4) and thence south to the preglacial valley near Covington;[13] (2) a possible "spur" connects Wabash and Illinois drainage;[14] and (3) the bedrock valley near LaFayette continues south through Fountain County, and the valley in Benton County represents an important western tributary.[15] Concerning the main valley at LaFayette, Fidlar further postulated[16] that this valley continued eastward into Ohio, where it joined the ancient Teays Valley[17] in the central part of the state near Chillicothe (Fig. 1). The course of the valley from LaFayette eastward across northern Indiana to the Indiana-Ohio state line in Adams County is based on numerous well records and in part follows the

[11] "Geology of Tippecanoe County," *Ind. Dept. Geol. and Nat. Res., 15th Ann. Rept.* (1886), p. 76.

[12] "Wells of Northern Indiana," *U.S. Geol. Surv. Water-Supply and Irrigation Paper 21* (1899), pp. 61–66.

[13] Leverett, p. 744 of ftn. 3. (1895).

[14] Clem, ftn. 4 (1911).

[15] M. M. Fidlar, "The Preglacial Teays Valley in Indiana," *Jour. Geol.*, Vol. LI (1943), p. 417.

[16] *Ibid.*, pp. 411–18.

[17] W. G. Tight, "Drainage Modifications in Southeastern Ohio and Adjacent Parts of West Virginia and Kentucky," *U.S. Geol. Surv. Prof. Paper 13* (1903), pp. 1–111.

course of a buried channel previously described by S. R. Capps.[18]

Near the Indiana-Ohio state line the channel coincides with preglacial channels discovered by J. A. Bownocker,[19] and the course southeastward to Chillicothe is based upon the work of Karl Ver Steeg.[20]

west boundary of Tippecanoe County and flows essentially on bedrock to a point about 3 miles south of Covington, where it again enters a buried bedrock valley. In this area Fidlar indicated that the preglacial valley followed a buried channel through Fountain County some distance east of the Wabash River and

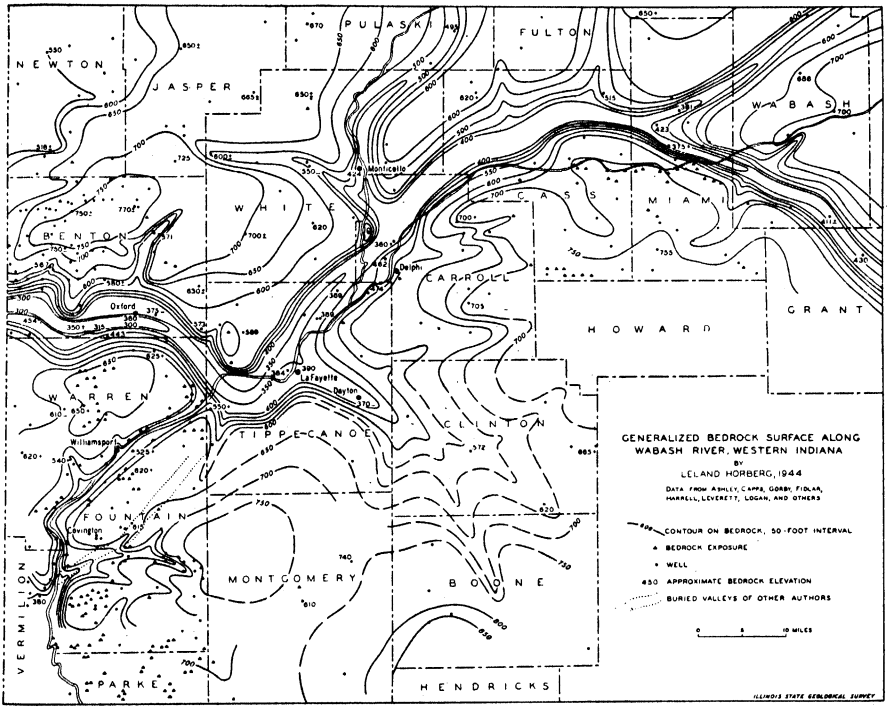


FIG. 4

The course of the ancient bedrock valley south of LaFayette does not follow the present Wabash River, as that stream enters a narrow valley near the

joined the present valley south of Covington (Fig. 4). As an alternative hypothesis it is proposed that the main valley turned west near LaFayette, through southern Benton County into Illinois and, as Mahomet Valley, continued west to join the bedrock valley along the Illinois River. The following considerations are offered in support of this interpretation: (1) the valley in southern Benton County appears to be comparable in

[18] "Underground Waters of North-central Indiana," *U.S. Geol. Surv. Water-Supply Paper 254* (1910), p. 26.

[19] "A Deep Preglacial Channel in Western Ohio and Eastern Indiana," *Amer. Geol.*, Vol. XXIII (1899), pp. 178–82.

[20] "The Buried Topography of Western Ohio," *Jour. Geol.*, Vol. XLIV (1936), pp. 918–39.

size to the valley above LaFayette, whereas any buried valley in Fountain County must necessarily be restricted in width; (2) bedrock elevations to the west are lower and more closely spaced than they are to the south, where there is an interval of about 90 miles before comparable low elevations are shown by Fidlar;[21] (3) this interval is an area of high bedrock indicated by well records and numerous bedrock exposures[22] (Fig. 4); (4) the published evidence supporting the buried valley through Fountain County is inconclusive,[23] and it is significant that Leverett[24] in a later discussion of the wells of Fountain County makes no reference to an important buried valley within the county.

No satisfactory interpretation of drainage changes in the area south of LaFayette can be made until further information becomes available. It is questionable whether the diversion from Mahomet Valley into the lower Wabash bedrock valley was directly into the present valley above Covington or through a buried valley to the east. Subsurface studies of the deposits filling the valleys are needed to establish the times of important erosion; and until this is done, interpretations will remain uncertain.

Although the details of drainage history in the LaFayette region are not entirely clear, the existing evidence strongly indicates that the main preglacial valley continued into Illinois as Mahomet Valley. If this is true and if the course of the ancient Teays east of LaFayette outlined by previous writers is confirmed, Mahomet Valley represents the course

TABLE 1

BEDROCK ELEVATIONS ALONG PROPOSED TEAYS VALLEY

Locality	Feet above Sea-Level	Reference
Seary, W.Va.	670	Stout and Lamb*
Chillicothe, Ohio	630	*Ibid.*
Madison County, Ohio	Less than 538	Ver Steeg, p. 925 of ftn. 20 (1936)
Jay County, Ind.	463	Fidlar, p. 416 of ftn. 15 (1943)
La Fontaine, Wabash County, Ind.	411±	Capps, p. 226 of ftn. 18 (1910)
Miami County, Ind.	423	Fidlar, p. 416 of ftn. 15 (1943)
Delphi, Carroll County, Ind.	360±	Logan†
LaFayette, Ind.	384	Fidlar, p. 416 of ftn. 15 (1943)
Oxford, Benton County, Ind.	300	Leverett, p. 757 of ftn. 3 (1895)
Rankin, Vermilion County, Ill.	358	Files, Ill. Geol. Surv.
Paxton, Ford County, Ill.	343	Savage, p. 444 of ftn. 5 (1931)
Mahomet, Champaign County, Ill.	357	Files, Ill. Geol. Surv.
Clinton, DeWitt County, Ill.	Less than 340	Files, Ill. Geol. Surv.
Delavan, Tazewell County, Ill.	288	Savage, p. 444 of ftn. 5 (1931)
Beardstown, Cass County, Ill.	311	Files, Ill. Geol. Surv.

* Wilber Stout and G. F. Lamb, "Physiographic Features of Southeastern Ohio," *Ohio Jour. Sci.*, Vol. XXXVIII (1938), also in *Geol. Surv. Ohio* ("Reprint Ser.," No. 1 [1939]), p. 14.

† W. N. Logan, "The Sub-surface strata of Indiana," *Ind. Div. Geol. Pub. No. 108* (1931), p. 47.

[21] Fig. 1, p. 412 of ftn. 15 (1943).

[22] Leverett, "Wells of Southern Indiana," *U.S. Geol. Surv. Water-Supply and Irrigation Paper 21* (1899), pp. 19–20.

[23] The evidence consists of a map and statement by R. T. Brown, published in 1881 without supporting data, in "Fountain County," *Ind. Rept. Geol. and Nat. Res. 11th Ann. Rept.* (1881), p. 92, map facing p. 89.

[24] P. 20 of ftn. 22 (1899).

of the lower Teays River.[25] By this hypothesis Mahomet Valley was eroded by a preglacial master-stream which probably had its source near the eastern scarp of the Blue Ridge in North Carolina;[26] flowed westward across Ohio, northern Indiana, and central Illinois; and finally discharged into the Gulf Embayment through bedrock valleys along the present Illinois and Mississippi rivers. Bedrock elevations (Table 1) along this valley indicate an average gradient of about 7 inches per mile for that portion of it above Beardstown, Illinois.

[25] In this case the name "Mahomet" should be dropped and the valley referred to as the "lower Teays."

[26] Wilbur Stout and Downs Schaaf, "Minford Silts of Southern Ohio," *Bull. Geol. Soc. Amer.*, Vol. XLII (1931), pp. 663–72.

ACKNOWLEDGMENTS.—This paper is an outgrowth of bedrock-surface and ground-water studies made at the Illinois State Geological Survey under the supervision of L. E. Workman, head of the Subsurface Division, and Dr. George E. Ekblaw, head of the Areal and Engineering Division. The early studies of Pleistocene stratigraphy in the area made by these two men assisted progress of the work, and the manuscript has benefited by their criticism. The aid and suggestions of other members of the Survey staff are also acknowledged.

REVIEWS

Handbook for Prospectors and Operators of Small Mines. By M. W. VON BERNEWITZ. Revised by HARRY C. CHELLSON. 4th ed. New York: McGraw-Hill Book Co., 1943. Pp. 547. $4.00.

During the fourteen years before his death, May 18, 1940, three editions had appeared of Mr. von Bernewitz' handbook. These editions not only were very useful to the mining industry but were opportunely issued so that the material was conveniently available through the entire third decade of the present century during the years of the great depression, when so many idle workers with little or no experience at prospecting were turning to it as a means of employment. Owing to the author's wide experience in North America, Australia, New Zealand, and the Netherlands Indies, as well as his journalistic experience, he was well qualified to undertake this work.

In the first three editions of the book von Bernewitz covered essentially the same ground except that in the second edition a chapter was added on geophysics. In the third edition this was withdrawn because it was regarded as too specialized, and it is not restored in the fourth edition.

Mr. von Bernewitz died before his fourth edition went to press. It was revised by Mr. H. C. Chellson, editor of the *Mining Congress Journal*, whose background is nearly similar to that of von Bernewitz, except that his mining experience has been chiefly in North America and Asia.

As the book now stands, the subjects are treated in six parts: (I) equipment, transportation, health, and laws relating to mining; (II) geology, sampling, and measurements; (III) metallic and nonmetallic minerals; (IV) ore dressing and treatment; (V) useful memoranda; and (VI) glossary.

Mining geologists will find (p. 66) a short and interesting list of examples of districts in which geology has proved useful in prospecting and in developing mines, with brief notes on some of them. Among those mentioned are the United Eastern, United Verde Extension, and Miami, Arizona; Chuquicamata, Chili; iron and copper mines in the Lake Superior district; gold mines in Western Australia. Many more could be added to this list.

The reviewer was particularly interested in notes on the Saun concession ("Colbran contact"), which lies near Holkol about 100 miles north of Keije (Seoul), the capital of Korea. The reviewer has not visited the mine but in 1921 was shown maps and other data of the district at the office of the geological survey of Chosen. This district, which was described by Higgins in *Economic Geology* in 1918 (XIII, 1–34), offers a fine example, showing how detailed geologic mapping aided by diligent use of the gold pan may assist in the discovery of a valuable deposit.

The book is compactly but strongly bound with light-weight flexible cover and is designed to stand the rough handling which books carried in the field are likely to get. The style is clear, and the material is well chosen. It contains little that will not be found useful for the inexperienced prospector and much that is conveniently arranged for the more experienced one. It is a very good book.

W. H. EMMONS

Climate of Indiana. By STEPHEN SARGENT VISHER. ("Indiana University Publications, Science Ser.," No. 13.) Bloomington, Indiana, 1944. Pp. 511; figs. 492. $4.00.

This is undoubtedly the most comprehensive study of the climate of a state which has appeared; and, as Indiana is representative of a considerable region, the findings here presented should also have important application well beyond the state boundaries. The volume contains an immense amount of factual material covering all aspects of Indiana's climate, much of which is made graphic on more than four hundred climatic maps of the state (mostly one-ninth of a page in size and generally arranged in groups of three, facilitating comparative study). It is a very thorough, painstaking piece of work.

Extending into the field of geology are two chapters: "Climatic Changes and Their Possible Causes" and "Climate and Physiography." These are subjects to which the author has given special attention for many years.

R. T. C.

VOLUME LIII NUMBER 6

THE JOURNAL OF GEOLOGY

November 1945

THE MOON'S LACK OF FOLDED RANGES

ROLLIN T. CHAMBERLIN
University of Chicago

ABSTRACT

The moon shows little evidence of folded mountain ranges. Faulting (especially in radial and concentric patterns) has been prevalent; and one area of aligned ridges, which extends several hundred miles from the Mare Imbrium, across the elevated environs of the Mare Serenitatis and the Mare Tranquilitatis, and is older than the flattish maria surfaces, shows some resemblance to our faulted Basin Ranges. But why are the long arcuate folded ranges, so characteristic of the earth, lacking on the moon?

What have been the notably different conditions on the moon and the earth which may have been responsible for their dissimilar diastrophic histories? Four differences come readily to mind: (1) The moon has neither atmosphere nor hydrosphere. (2) The moon has only 1/81 of the earth's mass, and its surface gravity is only about one-sixth as great. (3) Its average density is only 3.34, against 5.52 for the earth. (4) Its period of rotation is a little over 27 days, compared with the earth's daily rotation.

Some of the possible consequences of these differences are discussed briefly to indicate lines for further investigation.

INTRODUCTION

The most striking features on the face of the moon are the multitudinous craters, large and small, which scar its surface. These rimmed pits are generally thought to represent the moon's volcanic activities since very early in its history. An alternative hypothesis attributes them to explosions from the impact of large meteoritic masses. However formed, it is agreed that, in the absence of an atmosphere and hydrosphere, they escape the rapidly destructive gradational processes prevalent on the earth and should persist for great lengths of time, except as affected by later volcanism, meteoritic impacts, or lunar diastrophism. Much of the lunar facial expression is thus presumably very old.

But if the pits record topographic developments since remote times, where are the folded mountain ranges so characteristic of earth history? If formed, they likewise should be very long-lived. Photographs of the moon show surprisingly little evidence of folding. Some differences, of course, are to be anticipated. A corrugated lunar surface would remain uneroded; and we should not expect to see such close folding as in a deeply dissected terrestrial mountain system, where the folds commonly become sharper, smaller, and more numerous downward. Yet, if comparable mountain systems had developed on the moon, we should at least expect to see long belts of prominent, closely spaced wrinkles, with paralleling fault scarps; for rocks both bend and break, even when not subjected to confining pressures due to deep burial, as warped tombstones show. But on the moon we look in vain for great arcuate

strips of strong, paralleling corrugations, so familiar from our Appalachians, Juras, and many other terrestrial mountain systems.

LUNAR FAULTING

Faulting has occurred abundantly on the moon, although N. S. Shaler,[1] in spite of many years of study of photographs and by telescope, seems to have failed to recognize much of it as such. Many steep declivities in common alignment, however, can hardly be other than fault scarps and are now so regarded. Most characteristically these lunar faults fall into two categories: (1) a system radiating outward from a center of disturbance (like the Mare Imbrium); and (2) faults concentric with crater walls or with the margin of a mare (Fig. 1). Both these types show some resemblance to those developed on the earth by upswelling and collapse of domical uplifts. The first fault pattern may call to mind the array of dikes diverging from the Spanish Peaks intrusive mass in Colorado.[2] Upward pressure from underlying magma domes the surface, stretching and fracturing the rocks along radiating lines, which may become dikes or faults.[3] Radial fractures may also form because of horizontal pressure outward from a volcanic neck or a stock.[4] These may be added to those developed by doming. Faults of the second type, concentric with a lunar crater rim, step down its inner slope in successive blocks, somewhat as in the Kilauea caldera. Sinking of the caldera floor, or floor of a mare, by failure of underlying support, is the apparent reason for this succession of marginal normal faults. Common, also, is what looks like graben sinking between faults.

Recently a very thorough study and interpretation of the facial features of the moon, particularly as revealed in the Mare Imbrium region, has been presented by J. E. Spurr.[5] The Mare Imbrium, or Imbrian plain, is encircled by an elevated rim, greatly roughened by faulting, portions of which have been designated the "Apennines," "Carpathians," and "Alps." Along the inner side of these "ranges," as along other sections of the nearly circular rim, long fault scarps facing the mare show how that area has dropped from its surroundings in caldera fashion. But even more striking on Spurr's sketch map of the Imbrian fault system[6] are the very numerous transverse faults cutting the rim normal to the concentric downfaulting. Some of these are more than 100 miles in length. Converging, they point toward the middle of the mare; but they terminate abruptly at the mare border except in front of a portion of the Apennines, where lines of low scarps, peeping above the Imbrian plain, extend major transverse faults of the rim far out into the mare.

Subsequent to the faulting, according to Spurr, the sunken mare floor has been covered by vast floods of lava which obliterated its earlier relief and gave most

[1] "A Comparison of the Features of the Earth and the Moon," *Smithsonian Contributions to Knowledge*, Vol. XXXIV (1907), pp. 1–131.

[2] "Spanish Peaks Quadrangle," *U.S. Geol. Surv. Folio 71* (1901).

[3] T. A. Link (*Jour. Geol.*, Vol. XXXV [1927], pp. 327–39) has produced both ringlike and radial fractures by forcing a cylinder of plaster of Paris (and likewise liquid materials) upward into artificial strata. His concentric fractures, however, illustrate the formation of cone sheets better than the normal faulting around a sinking caldera.

[4] Willard H. Parsons, "Volcanic Centers of the Sunlight Area, Park County, Wyoming," *Jour. Geol.*, Vol. XLVII (1939), pp. 1–26.

[5] *Geology Applied to Selenology* (Lancaster, Pa.: Science Press Printing Co., 1944).

[6] *Ibid.*, p. 42, Fig. 11.

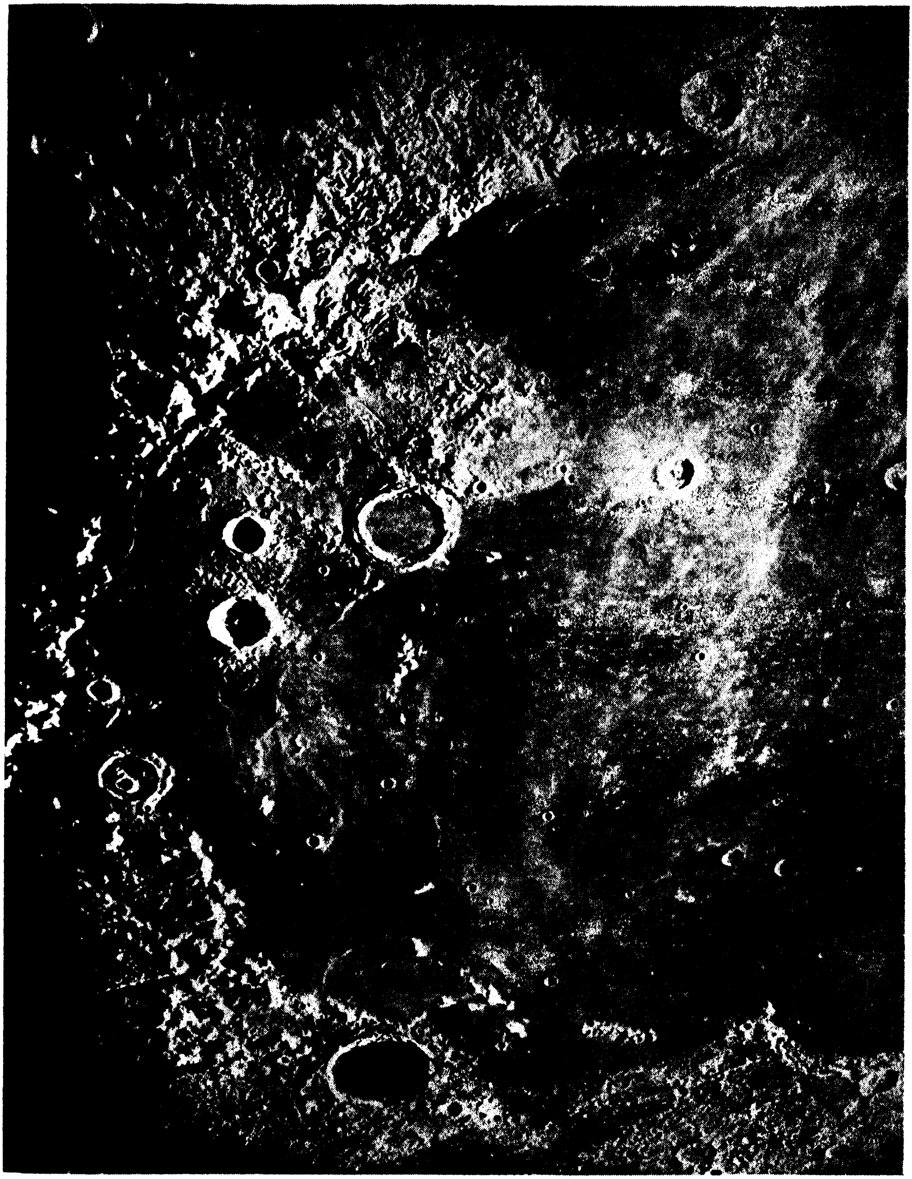


FIG. 1.—A portion of the Mare Imbrium and its elevated rim. The northwest section of the rim is termed the "Apennines." Faults scarps, both transverse to the rim (radial to the mare) and concentric with it, are evident. Photographed September 15, 1919, at the Mount Wilson Observatory, California.

of the mare its relatively flat surface.[7] But opposite the Apennines, it would seem that a patch of the older faulted surface remained too high to be buried completely, for here partially submerged scarps are still visible (see Fig. 1). Scarps concentric with the rim are likewise discernible.

The Mare Imbrium is over 700 miles in greatest diameter. How much of the faulting radiating from it may be attributed reasonably to tension during the postulated updoming and how much is better explained by horizontal outward shove from the mare area is uncertain. Horizontal movement along many of the transverse faults, subsequent to the concentric faulting, has been inferred by Spurr, whose diagrammatic sketches show much offsetting of the Apennine and Carpathian mountain fronts.[8] On the latter front, however, the younger mare surface reaches quite a few miles into the range between certain pairs of transverse faults, suggesting that these "bays" of the mare, or re-entrants in the range, occur where some fault blocks have dropped lower than their neighbors. In most of these cases the "sea" has penetrated considerably farther along one fault line than along the other, as if tilting of the included fault block had brought that side lower. Moreover, the "shoreline" appears too much and too variously curved to be explained solely as a basin border, offset by horizontal displacement on the faults (see Fig. 2). Differential vertical movement can readily explain most of the offsetting. How much horizontal slip may have occurred along these transverse faults, after the mare area began to sink and before the flooding, is consequently not readily determinable. But on the Apennine front, evidence of important strike slip on the radiating faults is more convincing. Compared, however, with the magnitude of the Imbrian mare, this has been rather moderate in amount.

While the pattern of faults diverging from the Imbrian mare is such that a genetic relationship seems likely, it is, nevertheless, surprising how far nearly parallel fault scarps extend northwestward[9] from the mare. The scarps are very discontinuous; but they have a common alignment for several hundred miles, producing a crude structural grain revealed by light and shadow (Fig. 3). They continue prominently across the elevated environs of the Mare Serenitatis and the near-by Mare Tranquilitatis.[10] In fact, these two maria have obliterated this older faulted surface over large areas and have done so without imparting any very noticeable fault pattern of their own. Whatever doming and collapse occurred in the development of these maria has left much less apparent effect on their surroundings than in the case of the greater Imbrian mare.

Perhaps, therefore, too much of this lunar faulting has been credited to the maria as by-products in their evolution. It may be noted that the most strongly developed radial faulting around the Imbrian mare is in this continuous tract of northwest-trending scarps, which show little genetic relationship to the Serenitatis and Tranquilitatis maria. Possibly this structural grain is the ex-

[7] Shaler, however, although favoring the volcanic hypothesis for the craters, considered extensive melting of the lunar surface by the impact of giant bolides the most acceptable explanation of the maria (pp. 15–21 of ftn. 1).

[8] Figs. 11 and 12 of ftn. 5.

[9] Photographs oriented with image inverted as seen through an inverting telescope. True orientation of the scarp trend would be southwestward as seen with naked eye or through field glasses, with north at top in northern hemisphere and west to the right.

[10] Pls. XVIII and XXII of ftn. 1.

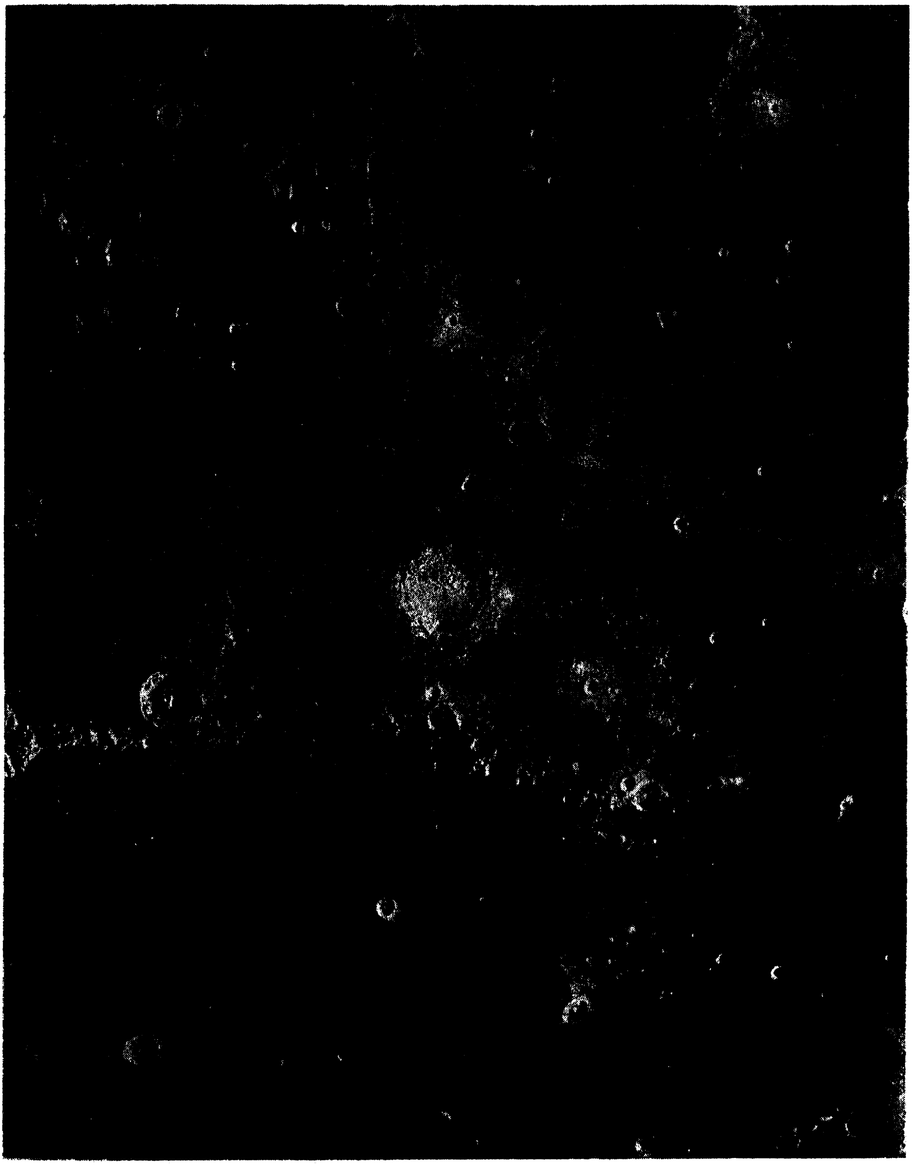


FIG. 2.—Region of Copernicus (central large crater). This photograph overlaps Fig. 1 and extends that area northeastward. The Imbrian mare occupies its lower third. The portion of the Imbrian rim just below Copernicus is termed the "Carpathian Range." Note the fault ridges transverse to this rim which radiate from the mare; also, the small "bays" of the mare which penetrate the rim. Photographed September 15, 1919, at the Mount Wilson Observatory, California.

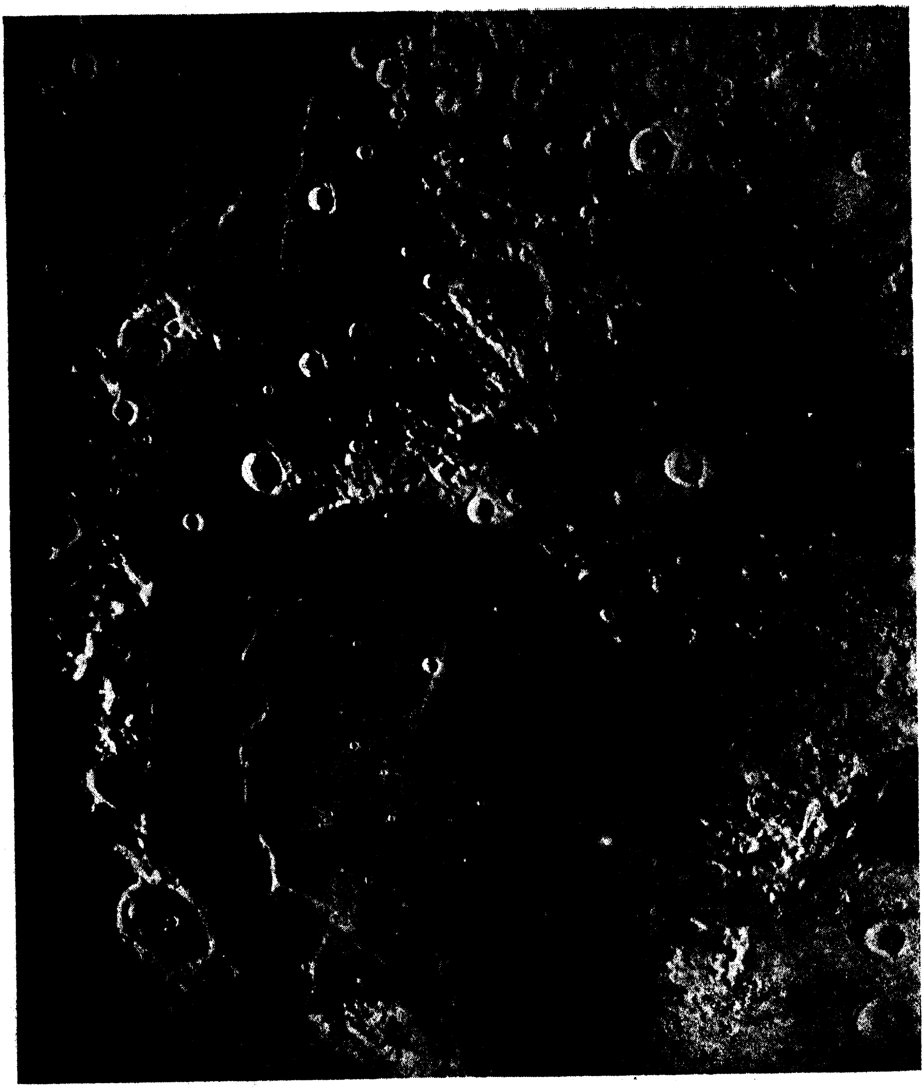


FIG. 3.—Mare Serenitatis (*lower center*) and Mare Tranquilitatis (*upper left*) with a bit of the Mare Imbrium in lower right corner. Note the northwest-southeast alignment of short, discontinuous ridges (presumably faulted) in the older uplands above the maria surfaces. Prominent serpentine wrinkle ridge runs north and south across the smooth Mare Serenitatis. Enlarged from a photograph made with the 40-inch refractor, Yerkes Observatory, Williams Bay, Wisconsin. Kindness of Dr. Otto Struve.

pression of a larger regional diastrophism (not altogether unlike the Basin Range faulting in the western United States), to which the Mare Imbrium has added its own fault system. Some contraction of the moon should be expected on any hypothesis of its origin, and the parallel faulting throughout this considerable tract may have resulted in part from such contraction. In any case, this faulting is older than the relatively smooth surfaces of the maria. Our only evidences of lunar diastrophism since the maria floors were formed are a few isolated faults and feeble wrinkles on their surfaces.

MARIA WRINKLE RIDGES

Some gentle wrinkling of the lunar surface is revealed by differences in illumination showing low, narrow upfolds, several of which are a hundred miles or more in length, whereas others are quite short. The greatest elevation assigned to any one of them, however, is less than 2,000 feet.[11] They are features of the maria floors and, although not numerous, are most prominent on the Mare Imbrium, the Mare Serenitatis, and the Mare Nectaris. Shaler states:

> So far as I have been able to ascertain, well developed continuous ridges are limited altogether to the maria and practically so to the larger fields of this nature; in the smaller maria they are much less distinct, though there are instances of slight undulations which may belong in the same category of structures.[11]

The most striking wrinkle ridge, which Shaler says "more distinctly resembles a terrestrial mountain chain than any other elevation on the moon," has a serpentine course across the Mare Serenitatis in a general north-and-south direction.[12] In some parts of its course (see Fig. 3) it is a single continuous ridge; in several places this gives way to two or more weaker, overlapping ridges; and toward each end of the chain the single ridge forks in widely diverging branches. Its over-all curvature is concave toward the middle of the mare. The echelon arrangement of individual ridges in several parts of this chain of arcs recalls strikingly the experiments of S. Tokuda,[13] who obtained similar fold patterns by placing a thin sheet of plastic paper coated with a thin layer of rice paste on a glass plate and then pushing the paper forward with his finger.

Spurr's analytical chart No. 1 of the Mare Imbrium[14] shows three main lines of wrinkling and various tiny ones with accordant orientation. Two lines curve around the western side of the mare (true orientation), while the other, at least 200 miles in length, bends around it on the east. Spurr says:

> This indicates that the mare flood, creeping up from the central part of the Mare, upon its "shores," was in its later stages plastic enough for the surface to be warped into wrinkles by the resistance of the bottom over which it advanced.

Certainly, no general lunar diastrophism is suggested by these few wrinklings of the maria floors in a late stage of their development.

How old the maria floors are, we do not know. But it is clear that, since their flattish surfaces were acquired, relatively little diastrophism has occurred on the moon. Meanwhile the earth has been producing great mountain systems from early pre-Cambrian times practically to the present.

[11] P. 36 of ftn. 1.

[12] *Ibid.*, p. 116, Pl. XVIII.

[13] "On the Echelon Structure of the Japanese Archipelagoes," *Japanese Jour. Geol. and Geog.*, Vol. V, Nos. 1–2 (1926–27), pp. 41–76.

[14] P. 7, Fig. 2, of ftn. 5.

DIFFERENT CONDITIONS ON MOON AND EARTH

The writer has long wondered at the lack of strong folding on the moon and has thought that its absence there should suggest leads toward a better understanding of terrestrial deformation. Spurr's recent application of geology to selenology has awakened dormant thoughts on the application of selenology to geology. Lunar conditions are different from terrestrial conditions, and in those differences one should seek the reasons why diastrophism has operated so dissimilarly on the two companion bodies. Presumably, therefore, the added vision from this new point of view may be brought to bear discriminatingly on the basic causes of earth deformation.

We ask ourselves, accordingly: What have been the notably different conditions on the moon and earth which may have been responsible for their dissimilar diastrophic histories? Four differences come readily to mind: (1) The moon has neither atmosphere nor hydrosphere. (2) The moon has only 1/81 of the mass of the earth, and its surface gravity is only about one-sixth as great. (3) Its average density is only 3.34, against 5.52 for the earth. (4) Presenting always the same face to the earth, it rotates but once during its $27\frac{1}{3}$-day circuit around the earth.

LACK OF ATMOSPHERE AND HYDROSPHERE

The ceaseless erosion of the earth's land surface by wind and water has no counterpart on the moon. These agents transfer no rock material from higher portions of the lunasphere to its lower surfaces, and no sedimentary rock formations develop. On the earth, folding into mountain ranges occurs principally where geosynclinal accumulation of weak sediments has reached many thousands of feet in thickness. Such accumulations are missing on the moon. In class discussions it has often sufficed to remark: "No geosynclines, no mountain-folding on the moon." But is this sweeping generalization really a complete answer? There can be little doubt that geosynclines have played an extremely important part in the great orogenies on our globe. They have been mobile belts of long duration; in general, they are favorably located with respect to growing earth stresses; weak sediments there extend to exceptional depths; and stratified sedimentary rocks fold more readily than nonstratified rocks by adjustments along bedding planes. That geosynclines should be the belts to yield, when tangential compressive stresses reach sufficient intensity, is natural. Strong folding takes place there more readily than elsewhere, and by this crustal shortening the stress conditions are relieved.

But while geosynclines locate the orogenies and unquestionably facilitate them, is it certain that their absence precludes notable folding? If we assume the same sources of tangential stress (whatever they are), would the earth's crust, unweakened by geosynclines, be sufficiently resistant not to shorten very considerably by buckling and thrust faulting? Trial computations based on assumed quantitative values might sharpen the focus on this question but probably would not resolve it, owing to the uncertain factors. However, it would not seem unreasonable to expect considerable crustal shortening by orogenic deformation without geosynclines, albeit less than with them and less concentrated in distinct belts.

Igneous rocks do not flex so readily as sedimentary strata, particularly near the earth's surface; but successive lava flows and pyroclastics do provide surfaces of weakness along which adjust-

ment can take place in bending, and even massive granite changes its shape without dependence on large-scale fracturing. In some of the great folds in the Rocky Mountains the smooth upper surface of the pre-Cambrian granite has a curvature identical with that of the basal Cambrian layer in contact with it. We say that the vertical or overturned Cambrian beds have been folded; but whether we call the bending of the granite-quartzite contact "folding" or not, the fact remains that the granite has deformed to the same configuration as the quartzite. Where the sedimentary formations have been stripped from the granite and other pre-Cambrian rocks, the latter now constitute the mountain ranges towering above the plains and near-by basins. Faults on an important scale may, or may not, be present.

Moon materials are less conducive to folding than those of the outer earth crust; but an orogeny by the requisite combination of flexing and faulting should be possible, provided the rocks are sufficiently stressed. That such has not occurred indicates insufficient tangential compressive forces. Absence of gradational processes eliminates one source of such forces. There is neither unloading of some areas by erosion nor loading of other areas by deposition of sediments.[15] Consequently, there would be no lateral crowding by a heavy subsiding segment against a lighter segment to operate, trigger-like, in setting off orogenic deformation by the general tangential forces in a shrinking globe. Likewise, much less isostatic adjustment would be brought into play. Also, without fluvial or eolian transfer of enormous masses of material from some portions of the moon's surface to other distant portions (as on the earth), there would be no shifting thereby of the moon's axis of rotation, with possible consequent distortion of the rotating globe. This last difference in diastrophic potentialities will be considered later.

[15] Ignoring very local talus.

LESSER MASS, GRAVITY, AND DENSITY

Although it is now rather widely recognized that the earth's crust has suffered far too much folding to be explained by shrinkage of the globe due to cooling alone, such contraction needs to be considered. An earth and a moon cooling from a postulated liquid condition to their present states should follow somewhat comparable courses. While the smaller moon would cool faster than the more massive earth and the ratios between the forces developed and the strength of materials would not be the same, the two bodies should undergo somewhat proportionate crustal shortening, with deformation likely to differ more in degree than in kind. But many ramifying factors would complicate any quantitative assay of the probable differences.

The small size of the moon may lead one quickly to the question: Assuming a completely molten moon, could it have undergone most of its cooling and contraction so early in its career that crustal shortening practically ceased before the maria were developed? They show little deformation; yet if we accept the volcanic theory for most of the lunar craters and maria, heat was still causing very extensive volcanism. Meanwhile the earth has been in orogenic throes, with little evidence of waning intensity, right to the present time. But we need not go further; cooling of the earth is no longer considered the principal cause of its crustal folding. To bring an estimate of the earth's crustal shortening into any sort of harmony with the total amount

obtainable under the thermal-contraction theory, it is necessary to ignore all the great pre-Cambrian folding, which is known to be intricate over vast surface areas and presumably also extends beneath millions of square miles of younger formations.[16] Also, mountain-building was very extensive in Cenozoic times, and we have no assurance that the earth was then losing much more heat than was being produced within it by radioactive processes.

Internal convection in a liquid state or even in solid material has, for several decades, been a postulated causal mechanism for orogenic deformation. In essentially solid globes, proximity of internal temperatures to the liquefaction point (represented by the temperature-pressure-fusion curve) should favor possible convective movements. Perhaps such a condition more nearly obtains in the earth than in the smaller satellite. But involved, also, would be liberation of new heat and unstable gravity relations. In the case of the moon, its strange ellipticity indicates that there has been no internal convection adequate to bring the satellite's shape into equilibrium with the forces operating during its later history.

The moon's weaker gravitation should affect deformation in several ways. Upbowing by tangential compression is opposed by less gravity on the moon, favoring the rise of anticlines and overthrust faulting. Fault blocks override with less friction. Cubic compression at a given depth below the surface strengthens the rocks less than at the same depth on the earth, though heat and other variables complicate the problem, which, indeed, is not a simple one. Larger blocks can remain out of isostatic equilibrium on the moon, and isostatic adjustment might be accomplished somewhat differently on the two globes. These and related differences are possible factors in comparative lunar and terrestrial diastrophism; but, obviously, they are not of the right sort to have been the dominating ones.

But gravity may operate more importantly in compaction and global shrinkage. One of the striking differences between the two bodies is a density of 5.52 for the earth and only 3.34 for the moon. The rocks near their surfaces should not differ radically in inherent specific gravity, but there may be considerable differences in porosity. Lavas should become more vesicular and pumiceous under low lunar pressures; and meteoric waters would not fill the cavities with mineral matter, although magmatic waters might. Accumulations of volcanic ejectamenta could retain high porosity to considerable depths. But such would be offset by porous terrestrial sediments and the low density of our oceans.

The excess of 5.52 over 3.34 comes chiefly from heavier material deep in the earth. Regardless of a postulated greater proportion of iron in the earth than in the moon, much of the earth's higher density results from the greater compression of the larger body. E. D. Williamson and L. H. Adams[17] have given curves for the change of density due to compression alone from the earth's surface halfway to its center. For two initial densities, 3.0 and 3.5 (corresponding to average gabbro and dense peridotite), their curves rise to 4.9 and 5.5, respectively, at a depth of 3,200 km. They indicate very considerable compression, although "it is clearly evident that the density is not increasing fast enough to make the

[16] See *Jour. Geol.*, Vol. XXXII (1924), p. 550.

[17] "Density Distribution in the Earth," *Jour. Wash. Acad. Sci.*, Vol. XIII (1923), p. 419, Fig. 2

mean density of the earth equal to 5.5"; to obtain this high figure, we must fall back on the only reasonable alternative, namely, that the deep interior is of intrinsically heavier material, probably iron.[18] In any case, it is clear that in self-compression lies one of the notable differences between earth and moon.

Many years ago, T. C. Chamberlin urged the importance of rearrangements in the interior of the earth in favor of greater density as a means of earth contraction and cause of megadiastrophism. This was a contraction not dependent on earth-cooling, whose inadequacy to produce the geologic results was becoming apparent. Self-compression of the globe involves far more contraction than is possible from cooling; but it must, of course, have operated through the span of geologic time if it is to meet the requirements of the earth's diastrophic history. Simple mechanical compression would not be so protracted a process; but appeal was made to subsidiary processes such as recrystallization, or neocrystallization, through slow molecular transfer, forming denser combinations of smaller volume under enormous pressures and varying thermal conditions. Conceivably, such a process might be very slow and intermittent in operation.

A. N. Winchell[19] has shown that, in regions of extensive and repeated intrusions, several effective factors operate, through selective fusion and solidification, to segregate the lighter, more acidic constituents of the lithosphere above and the heavier, more basic constituents below. If there were no shell-like arrangement initially, such would develop. Geszti has investigated the problem of gravimetric differentiation at greater depths. Beno Gutenberg epitomizes some of his results in the following words.

> As, in the deeper parts of the earth, gravity decreases with depth reaching zero in the center, all processes that are controlled by gravity must decrease in speed toward the center of the earth. Relatively small accumulations of materials that are heavier than their neighborhood will not sink toward the center of the earth. Only after larger masses of heavier materials have accumulated may extended mass movements occur, which will produce, in turn, great disturbances at the surface. Geszti correlates such revolutions in the earth with the geological epochs in which large changes have occurred.[20]

Much metallic iron is postulated in the central portion of the earth, but the low density of the moon will not allow so much heavy iron in its constitution. Because of this difference and because of the feebler lunar gravitation, the moon should not have the potentialities of the earth for internal rearrangement. Other factors may enter into the problem. P. W. Bridgman's researches[21] on the effects of high pressures have yielded some surprising results, and we cannot dismiss the likelihood of more surprises in store for us could the conditions deep in the earth be studied experimentally. Polymorphic transitions are, at least, suggestive of one line of possibilities. Changes of state and convection movements from radioactive heat have been variously considered in the problem of earth deformation.

SLOWER ROTATION

The moon rotates but once while the earth spins around twenty-seven

[18] *Ibid.*, p. 420.

[19] "Are There Granitic and Basaltic Shells in the Earth?" *Science*, Vol. LXXIX (1934), pp. 32–33.

[20] *Physics of the Earth*, Vol. VII: *Internal Constitution of the Earth* (New York: McGraw-Hill Book Co., 1939), p. 183.

[21] "Shearing Phenomena at High Pressure of Possible Importance for Geology," *Jour. Geol.*, Vol. XLIV (1936), pp. 653–69; "Polymorphic Transitions and Geological Phenomena," *Amer. Jour. Sci.*, Vol. CCXLIII-A (1945), pp. 90–97.

times. This is one of the striking differences between the two bodies. If enough of the earth's great rotational force should become available from time to time for tangential orogenic compression, herein might lie a reason for abundant folded mountain ranges on the earth and their scarcity on the moon. The lunar case being negative, does the earth offer any encouragement?

Changes in the speed of earth rotation have often been considered but have not proved very helpful. Shifting poles has been such a convenient way of explaining past climates and solving other geologic difficulties that possible forces to produce such shiftings have received considerable mathematical attention. Sir George Darwin[22] considered changes in the shape of the globe by erosion, deposition, and especially elevation and subsidence of large areas. He estimated that, if 1/200 of the earth's surface (centered in latitude 45°) be elevated 10,000 feet (effective elevation), the pole of rotation will be deflected $11\frac{1}{8}'$; if 1/20, $1° 46\frac{1}{2}'$; if 1/10, 3° 17′. In each case an equal area is supposed to fall simultaneously. Darwin's figures apply to an unyielding earth; a yielding earth would require greater displacement of matter to effect the same displacement of the axis of rotation; but both Darwin and Schiaparelli fell into error in handling the case of a mobile earth.[23]

Gutenberg gives a concise treatment of the many forces connected with the rotation of the earth that have been investigated.[24] He offers little encouragement for the hope that any of them will be found important in mountain-making. At the same time, he does state:

> Larger strains are found if we suppose the existence of extended movements of the earth's axis during geologic eras. A displacement of the poles by even a few degrees requires maximum vertical movements exceeding 1 km. in order to restore the equilibrium figure of the earth. These movements, of course, do not produce mountains but have a tendency to shift the diameters of the earth into new positions. However, large strains would be connected with such an event, and, if great displacements of the poles occurred during the history of the earth, large strains would have been a consequence, especially in the meridian along which the poles were traveling.

A year ago, Mr. Joel E. Fisher published privately a small book, entitled *Problems in the Geology of Mountains*, as a scientific by-product of many seasons of high-mountain climbing. Resulting from more intimate and extended contact with névés and the upper reaches of glaciers than most professional geologists possess, his three chapters on *nieves penitentes*, Forbes dirt bands, and the origin of glacial cirques were a notable contribution to glacial geology, and the present writer reviewed the book for the *American Alpine Journal*.[25] A fourth chapter, "Shiftings of the Axis of Gyration of the Earth as an Originating Mountain-making Force," presented the thesis that extensive redistribution of earth mass—whether by erosion, deposition, accumulation of ice caps, uplifts, subsidences, or movements of magma—would shift the earth's axis of rotation and that this, in turn, would lead to the folding and uplift of mountain ranges. The prime question being the magnitude of the possible tangential forces, which are generally considered too small to be important in mountain-making, the reviewer suggested making quantitative tests of selected mountain systems by a

[22] G. H. Darwin, "On the Influence of Geological Changes on the Earth's Axis of Rotation," *Amer. Jour. Sci.*, Vol. XIII (1877), pp. 444–48.

[23] W. D. Lambert, Frank Schlesinger, and E. W. Brown, "The Variation in Latitude," *Physics of the Earth*, Vol. II: *The Figure of the Earth* (1931), pp. 266–67.

[24] Pp. 166–74 of ftn. 20.

[25] Vol. V, No. 3 (1945), pp. 422–25.

series of calculations based on reasonable assumptions of how much earth mass might be displaced by recognized geologic processes.

This Mr. Fisher has undertaken to do; and in an article to appear before this,[26] he presents among his results the estimate that a tilting of the earth's axis of gyration by 10′ may be expected to produce uplifts of the order of magnitude of the Alps-Caucasus-Himalaya, or the American Cordillera, and that the transportation of the widespread sediments of Cretaceous time may well have caused a 10′ tilt of the earth's axis. Still, one may suspect that during the slow transfer of this great bulk of sediment, isostatic and other internal adjustments might limit the amount of axis-tilting much more than the author has thought; but quantitative evaluation is not easy. On the whole, his method of testing, by calculating the change in bulge-lifting force, for various latitudes, consequent on a 10′ tilt of the axis of rotation, has not yet shown convincingly that there has been developed thereby sufficient intensity of tangential compressive stress in the outer earth crust to corrugate a geosyncline into a mountain system. Further analysis of the problem is highly desirable because, if Fisher is right, the results are of great importance.

If rotational forces should prove more important in orogenesis than heretofore thought possible, they would be a factor for consideration in the problem of lunar versus terrestrial diastrophism. On the smaller, slow-turning moon they are much weaker than on the earth. Transfer of lunar material by atmosphere and hydrosphere is lacking, thus eliminating that possible cause of axis tilting. Also, the ellipticity of the moon, instead of being adjusted to the satellite's present distance from the earth and present speed of rotation, is about what it should be were the moon only one-third as far from the earth as now, and rotating faster, which indicates that the moon has sufficient strength to maintain for a long period of time the shape acquired presumably under a different rotation and much stronger earth pull.[27] Forces connected with the moon's rotation should not be expected to fold its surface into mountain ranges.

CONCLUSION

Diastrophism has not affected the surface of the moon in the same way, or to the same extent, as the earth. The dissimilar behavior must have resulted from different governing conditions. The most obvious differences between the two bodies which might have been responsible for the very different diastrophic manifestations have been considered briefly in reconnaissance fashion. The differences are great, but each ramifies into various complicating factors which must be understood and evaluated in reaching any firm conclusions. The general problem must be broken into its component parts, each of which will require careful study. But, in any case, since lunar diastrophism has been the product of lunar conditions, as terrestrial diastrophism has been the product of earth conditions, comparative studies of moon and earth will enable us to survey certain aspects of the earth's deformation with the added perspective of another point of view.

ACKNOWLEDGMENT.—The writer wishes to express his indebtedness to Dr. Otto Struve, of the Yerkes Observatory, who kindly supplied many excellent photographs of the moon, only three of which are here reproduced as illustrations.

[26] "A Kinetic Theory on the Origin of Orogenic Forces," *Amer. Jour. Sci.*, Vol. CCXLIII (1945), pp. 606–13.

[27] Pp. 37–38 of ftn. 20.

THE PITTSBURGH-POTTSVILLE BOUNDARY

GEORGE H. ASHLEY
State Geologist of Pennsylvania

ABSTRACT

The base of the Brookville coal (or the top of the underlying Homewood sandstone) has long been the accepted Pittsburgh-Pottsville (Upper-Lower Pennsylvanian) boundary in the Appalachian bituminous coal fields. But increasingly for the last forty-five years, detailed studies have shown that the boundary has been drawn at many horizons, some over 500 feet apart in the same section, owing to differing correlations. The reasons for these diverse correlations are given; and as the simplest solution of the problem, it is proposed to shift the boundary to the base of the Lower Kittanning—No. 5 Block coal, believed to correlate with the Buck Mountain bed, whose base forms the boundary at the type Pottsville locality at Pottsville, Pennsylvania.

THE BROADER PROBLEM

The Pennsylvanian system of the Appalachian region has from the beginning of its study been divided into two series. The upper, which contains the Monongahela, Conemaugh, and Allegheny groups, was called the "Coal Measures" in the early literature, or more recently the "Pittsburgh series."[1] The lower series, at first known as the "Conglomerate," has long been called the "Pottsville." The line between these series was drawn originally at the base of "the lowest coal seam"; and it was assumed that this bed was at the same horizon in the anthracite, Broad Top, and bituminous areas. This much appears in the *First Annual Report* of the Rogers Survey in 1836 (pp. 16–17). Today, owing to a number of causes, the boundary has been drawn at many horizons, some as much as 500 feet apart in the same section. This paper undertakes to resolve these differences and to present a satisfactory, uniformly defined boundary.

DISCREPANCIES IN PLACEMENT OF BOUNDARY

In 1900 David White[2] pointed out that the top of the Pottsville, as then drawn in the Northern Anthracite field and in the bituminous fields, did not correspond with the boundary as drawn at Pottsville.

> According to the evidence of the plants, the beds of the basal portion of the Allegheny series —as high as the Clarion coal in the bituminous basins of western Pennsylvania—and the terraines extending from Coal A to at least as far as Coal C in the Northern Anthracite field, were laid down prior to the deposition of the roof shales of the Twin (Buck Mountain) coal,

which forms the boundary in the type locality.

Time has shown that the problem and practice is much more complicated than is indicated by White's statement. The irregularities of the beds and the lack of persistent key rocks and of short-lived plant and animal species, by which to distinguish one horizon from another, have resulted in the conditions shown graphically in Figure 1.

The boundary has been placed at different horizons for four reasons:

1. As pointed out by David White, errors all due to assuming that the lowest workable coal bed in the several fields correlates with the Buck Mountain bed

[1] G. H. Ashley, "A Geologic Time Scale," *Eng. and Min. Jour. Press*, Vol. CXV (1923), pp, 1106–8.

[2] "Stratigraphic Succession of Fossil Floras of the Pottsville Formation in the Southern Anthracite Coal Field, Penna." *U.S. Geol. Surv. 20th Ann. Rept.* (1900), Part II, pp. 829–30.

at Pottsville; or, quite as often, that the massive sandstone underlying the minable coal beds is at the same horizon as the massive sandstone underlying the Buck Mountain bed at Pottsville. This has led to using the top of the Homewood (secs. 3, 5, 6) of the upper Connoquenessing (secs. 3–6), and of the Roaring Creek sandstone (secs. 13–17) at three different horizons, none of which correlates with the massive sandstone at the top of the Pottsville at Pottsville.

2. The extreme irregularity of the beds lying below the level of the Buck Mountain–Lower Kittanning coal has made it difficult to trace these beds from county to county or often from one valley to the next. This has caused many local variations in the mapping of these beds. Then, as discussed later, detailed mapping has shown that the Brookville coal of the middle Allegheny Valley is the Upper Mercer coal of Beaver Valley, 40–50 feet or more below the coal mapped as Brookville in the Beaver Valley.

Figures 2, 3, and 4 reveal in graphic form the irregularity of the Mercer coals ("Brookville" of recent folios in Allegheny Valley). The sections of Figure 2 were measured by Charles K. Graeber in the new Humphrey clay quarry $1\frac{1}{2}$ miles southeast from the center of Brookville. Figure 3 shows sections taken near Burly, in southeastern Clearfield County. These show the tendency of at least one of the Mercer coals to divide or run together or disappear. Figure 4, sections near the Pennsylvanian Railroad station at Clearfield, shows also the variability of the A (Brookville) coal within a short distance.

3. The slow vertical changes in the fauna or flora over large intervals have made difficult the correlation of beds, from point to point, by fossils. Thus, a study of the Pottsville fauna in Ohio by Helen Morningstar[3] showed that (see sec. 10), of 35 species in the Lowellsville limestone below the Lower Mercer, all but 6 occur in the Lower Mercer. Of 65 species in the Upper Mercer clay bed, all but 7 are in the Lower Mercer. Of 94 in the MacArthur limestone (equal to Putman Hill, according to Wanless), all but 9 are common with the Lower Mercer; and of 52 in the Black Flint, which is today placed above the Putman Hill, all but 3 are found in the Lower Mercer horizon. Likewise, David White pointed out that it was easy to distinguish the differences in the plants at the Kittanning horizon or below the Lower Mercer horizon, but not between the intermediate horizons. White depended largely on minute differences too small to serve even as a basis for subspecies to distinguish one flora from another. The writer is one of many who put much weight on White's age determinations because of his long experience and close study of the Pennsylvanian plants. Thus, in a letter to David Reger of July 5, 1929, White states: "1. Coal No. 4 of Carter and Boyd Counties, Kentucky, is Upper Mercer (Section 12). 2. So is the Lower Torchlight coal without possibility of question. 3. Zaleski flint I have regarded as Kanawha black flint, = Mercer." See sections 12–16. If White is right, then the Winifred coal of West Virginia cannot possibly correlate with the Brookville coal of Ohio or northwestern Pennsylvania, and the Winters and Ogan coals = Upper and Lower Mercer and the "Clarion coal" (sec. 10) are the Brookville (Fig. 5). See also Morningstar,[4] who draws the top of the Pottsville at the base of the coal above the black flint

[3] "Pottsville Fauna of Ohio," *Ohio Geol. Surv. Bull. 25* (4th ser., 1922), pp. 138–44.

[4] *Ibid.*, pp. 134–36.

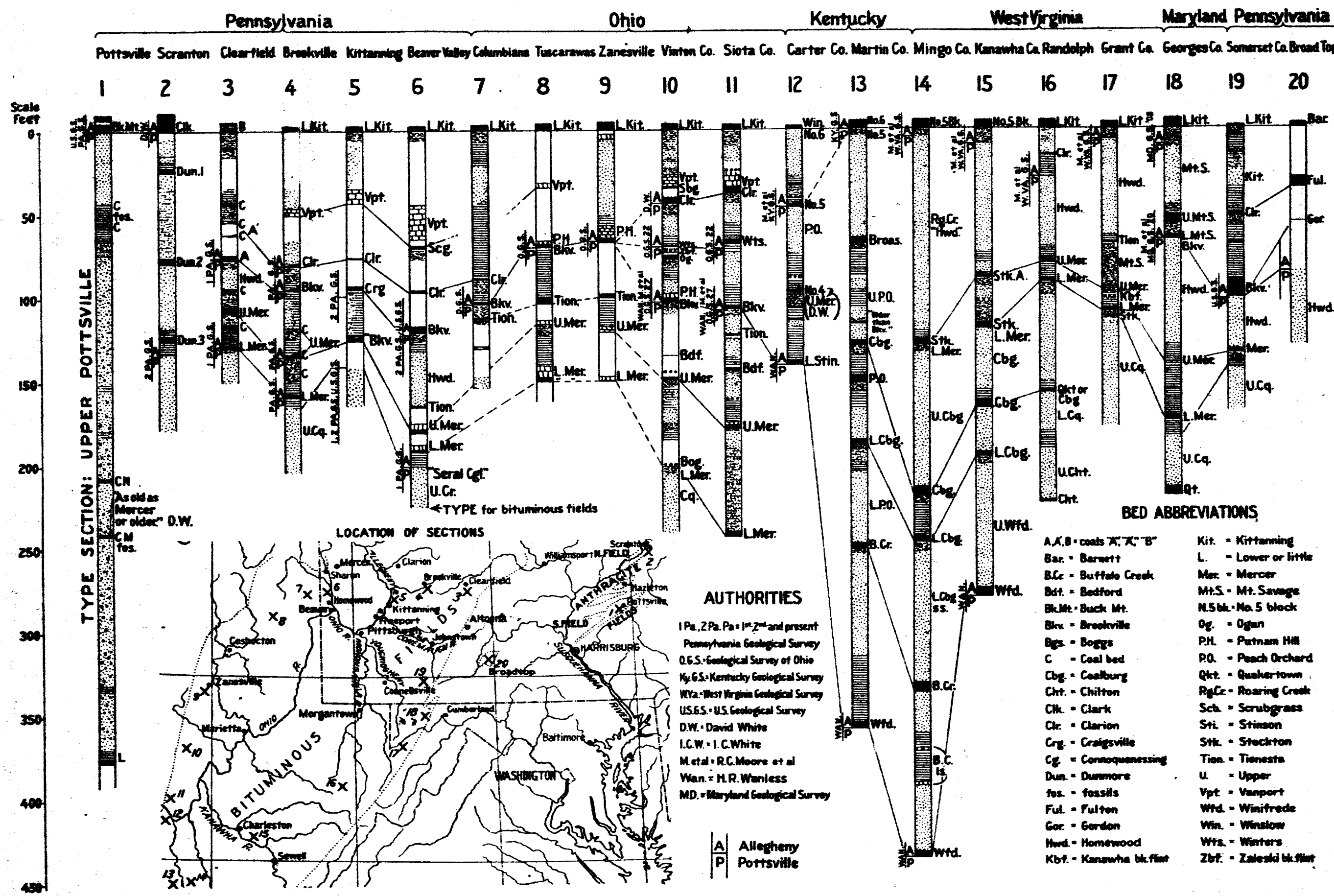


FIG. 1.—See legend on facing page

(secs. 10 and 11). The Kentucky and West Virginia geological surveys (secs. 12–17) have accepted and use White's correlations, claiming that they are in accord with their own tracings.[5] These instances are cited simply as evidences of the difficulties encountered in attempting to make correlations by means of fossils. Wanless[6] used the most modern methods except detailed field tracing.

4. On the other hand, as shown in Figure 5, if, as correlated by White, the Sharon coal of Pennsylvania, 250–300 feet below the Lower Kittanning, is the stratigraphic equivalent of the Sewell or the Iaeger coals of southern West Virginia, 1,900–2,300 feet below the No. 5 Block, it would seem more reasonable to consider that the entire northwestern Pennsylvania section had been thickened in West Virginia rather than that the lower part was tremendously thickened while the upper part had been thinned, as would be necessary if the Roaring Creek sandstone correlates with the Homewood and the Stockton coal is the same as the Lower Mercer. This would favor Wanless' correlation over the others.

[5] D. B. Reger, *County Reports: Mineral and Grant Counties, W. Va. Geol. Surv.* (1924), pp. 250–52.

[6] "Pennsylvanian Correlations in the Eastern Interior and Appalachian Fields," *Geol Soc. Amer. Special Paper 17* (1939).

FIG. 1.—Sections to bring out the varied usage in drawing the boundary between the Pittsburgh and Pottsville series or between the Allegheny and Pottsville. The various boundaries in use are indicated by the *A* over *P* at the left side of each section. This varied practice has led to the belief that a solution is needed.

Location and source of sections:

1. Pottsville, Pa., Southern Anthracite field, D. White, *U.S. Geol. Surv. 20th Ann. Rept. of the Director*, Part II, Pl. CLXXXI.
2. Scranton, Northern Anthracite field, *2d Geol. Surv. of Pa.*, Vol. AA: *Northern Anthracite Atlas III*, Columnar Section Sheet No. IX, Sec. V. Lower boundary by Second Geol. Surv.; upper, by D. White.
3. Clearfield, G. H. Ashley, *Pa. Geol. Surv. Bull. M6* (1932), Part II, p. 158. Lower boundary by Rogers; upper, by Second Geol. Surv. and later.
4. Brookville (comp.). See *Pa. Geol. Surv. Atlas A54* (1942), p. 62, Brookville quadrangle. (See discussion.)
5. Kittanning, Charles Butts, *U.S. Geol. Surv. Bull. 276* (1906), Pl. V, pp. 32–35. Intervals checked by measuring intervals on the map and averaging. Lower by First and Second Pa. Surv. and U.S. Geol. Surv.; upper by Second Surv. by miscorrelation.
6. Beaver Valley (comp.). See *Pa. Geol. Surv. Atlas 3*(1929), pp. 35 and 47, New Castle quadrangle.
7. Columbiana County, Ohio, W. Stout and R. E. Lamborn, *Ohio Geol. Surv. Bull. 28* (4th ser.; 1924), p. 51.
8. Tuscarawas County, Ohio, E. Orton, *Ohio Geol. Surv.*, Vol. V: *Economic Geology* (1884), p. 67.
9. Zanesville, Ohio, E. Orton. See p. 96 of ftn. 14.
10. Vinton County, Ohio, W. Stout, *Ohio Geol. Surv. Bull. 31* (4th ser.; 1927), pp. 66, 67, and 160.
11. Sciota County, Ohio, H. R. Wanless, *Geol. Surv. America Special Paper No. 17* (1937), Pl. IV, sec. 3.
12. Carter County, Ky., H. R. Wanless (see ftn. 17), Pl. IV, sec. 6 (lower boundary). D. White identifies Ky. No. 4 coal as Upper Mercer (see discussion). No. 5 coal, therefore, must be the Brookville. The Ky. Geol. Surv. has adopted that boundary in contrast with Wanless' use of the Lower Stinson. The same is true for Martin County and those following.
13. Martin County, Ky., Wanless (see ftn. 17), Pl. IV, sec. 9.
14. Mingo County, W.Va., Wanless (see ftn. 17), Pl. IV, sec. 18.
15. Kanawha County, W.Va., Wanless (see ftn. 17), Pl. IV, sec. 22.
16. Randolph County, W.Va., *W.Va. Geol. Surv. County Rept. 1931*, pp. 223, 232, and 233.
17. Grant County, W.Va., D. B. Reger, *W.Va. Geol. Surv. County Rept. 1924*, pp. 184–87, Mineral and Grant counties.
18. Georges Creek Basin, Md., C. K. Swartz, *Md. Geol. Surv.*, Vol. XI (1922), pp. 125 and 126. Note change from report of same survey in 1905.
19. Somerset County, Pa., G. B. Richardson, *U.S. Geol. Surv. Fol. 224* (1934), Somerset-Windber quadrangle.
20. Broad Top Field, J. H. Gardner, *Topog. and Geol. Surv. Comm. Rept. 10* (1913), p. 26.

PROPOSED DEFINITION OF BOUNDARY

I propose that the base of the clay under the Buck Mountain coal and of

other coals correlated with it be made the accepted plane between the Pittsburgh and the Pottsville series.

Therefore, it is necessary to justify the recommended change. The change is proposed solely because of the past and

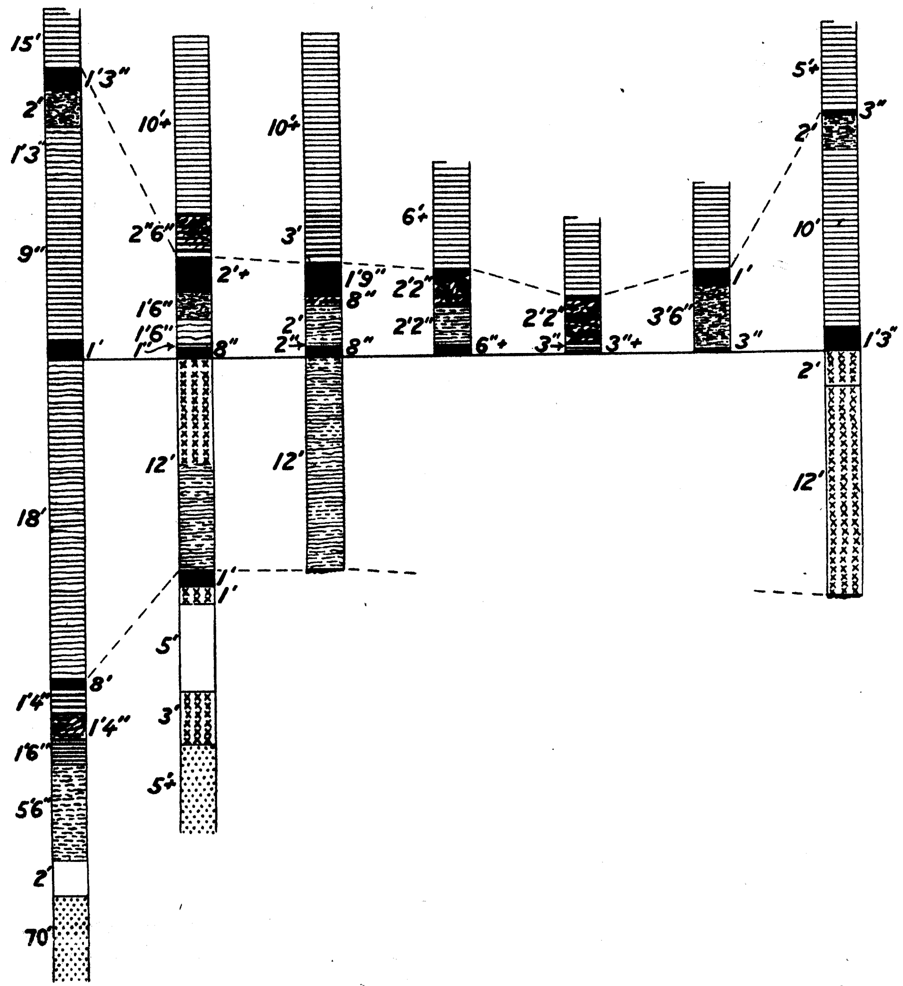


FIG. 2.—Measurements on Mercer coals in face of New Humphrey quarry, $1\frac{1}{2}$ miles southeast of Brookville (Graeber).

Throughout the bituminous fields the base of the "Brookville coal" or the top of the underlying "Homewood sandstone" has been widely used as the top of the Pottsville. The proposed definition departs completely from that practice.

present variations in usage and because there seems to be no way of resolving those differences except by a detailed and painstaking tracing of the beds in the field, supplemented by considerable drilling data. Such a survey may be a long

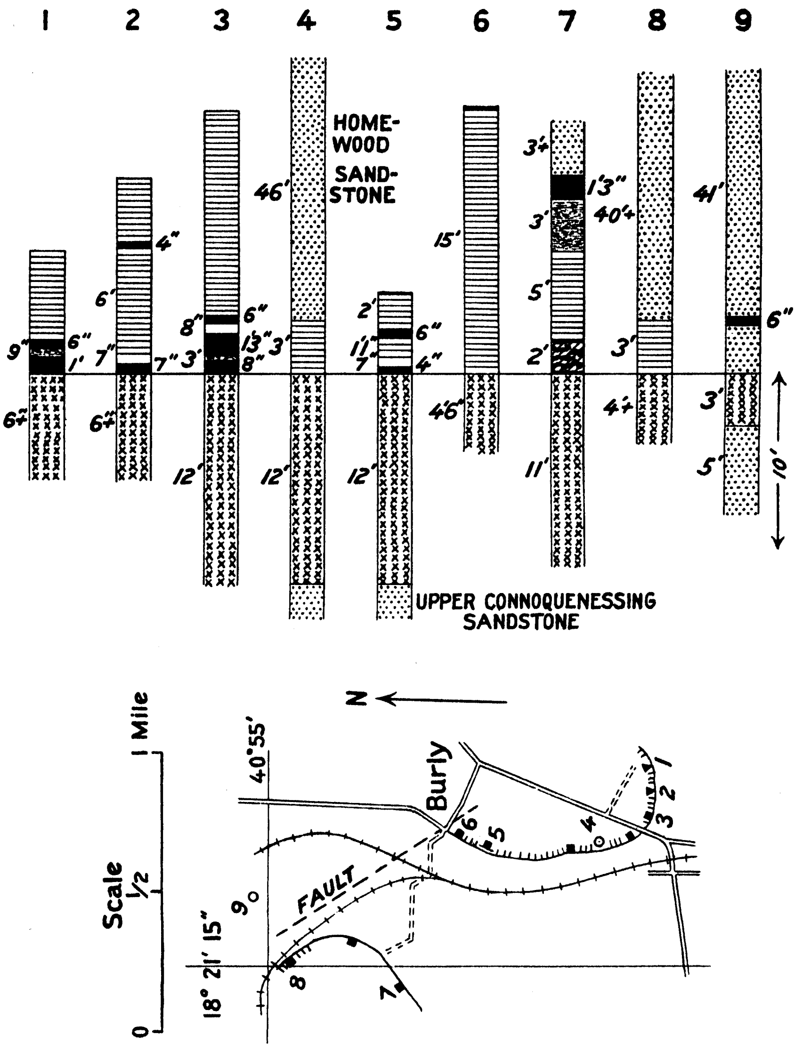


FIG. 3.—Measurements on Mercer coals in strippings and drillings near Burly in southeastern Clearfield County. These are exposed only because of the commercial value of the refractory clay shown in sections.

time in the future. The discussion which follows seeks to trace the history of the varied boundary definitions and to present all of the facts.

DISCUSSION

FIRST GEOLOGICAL SURVEY OF PENNSYLVANIA

In his *First Annual Report* (1836, p. 17) Rogers included in the Coal Measures all the beds "above the red shale strata, notwithstanding that the lowest rocks are destitute of coal through a thickness of several hundred feet." In the *Second Annual Report* (1837) Rogers presented a table, from which the following is taken:

TABLE OF STRATIFICATION (IN PART)*

		Feet
XIII	"Shales, sandstones, and seams of Anthracite coal"	6,750
XII	"Coarse quartzose conglomerate," etc.	1,400
XI	"Red shales and soft argillaceous red sandstones"	2,949

* Rogers, *2d Ann. Rept.* (1837), opp. p. 19.

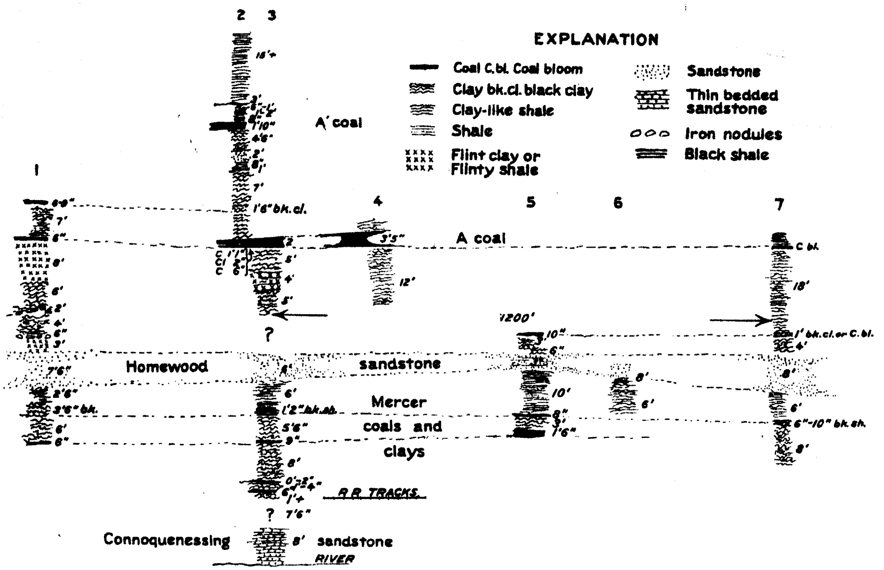


FIG. 4.—Exposures of coal and clay at Mercer and Brookville horizons, just east of Pennsylvania Railroad station at Clearfield.

Formation XII is described as "subjacent to all of the coal measures of the State" and "reposes on the red shale." The description of Formation XIII is devoted to certain structural details and does not specifically mention the individual coal beds. Nowhere in the annual reports of the First Survey is the upper limit of the conglomerate in the anthracite fields clearly defined, probably owing to the crushed condition of the coals, except to state that it included everything up to the base of the "first coal" of the coal measures. By "first coal" was meant "the lowest workable coal."

In the bituminous fields the line between formations XII and XIII was first drawn in Rogers' *Fourth Annual Report* (1840) (p. 177), which describes the "Coal Measures—Allegeheny Series" along Allegheny River northward from Pittsburgh, in descending stratigraphic order. "No. 6—Coal" is the "first workable coal seam of the Allegheny series"

and is clearly the Upper Freeport of today; "No. 12—Coal" is Lower Kittanning; "No. 15—Fossiliferous Limestone" is the Vanport; "No. 17—Coal" is the Clarion; "No. 18—Shale and argillaceous sandstone" "about forty feet" thick, overlies "No. 19—Coal"; and beneath this is "Formation XII—White Sandstone." Number 19 is "the lowest workable coal hitherto found in the series along the Allegheny River." This seems clear enough; since "No. 17—Coal" was 20 feet below the Fossiliferous limestone (Vanport) and "No. 19—Coal" 40 feet lower, the top of Formation XII should be about 60 feet below the limestone or in the position of the top of the Tionesta sandstone as given in Rogers' 1858 section. But his description of "No. 19—Coal" fits the Upper Mercer coal of today, for it crops out "in a northeast–southwest direction through Mercer, Venango, and Clarion Counties." Specific locations given are all in Venango County, and they obviously refer to the Upper Mercer coal as now identified in that area. But the Upper Mercer coal averages 100 feet or more below the Vanport limestone in that area, instead of 55 feet, as indicated in the Rogers' section given below.

In the final report of 1858 Rogers gives the following compiled section of his basal Allegheny series.

This quoted section leaves no doubt that in the end Rogers drew the line between formations XII and XIII in the bituminous coal fields at the base of the Mercer member; nor is there any doubt

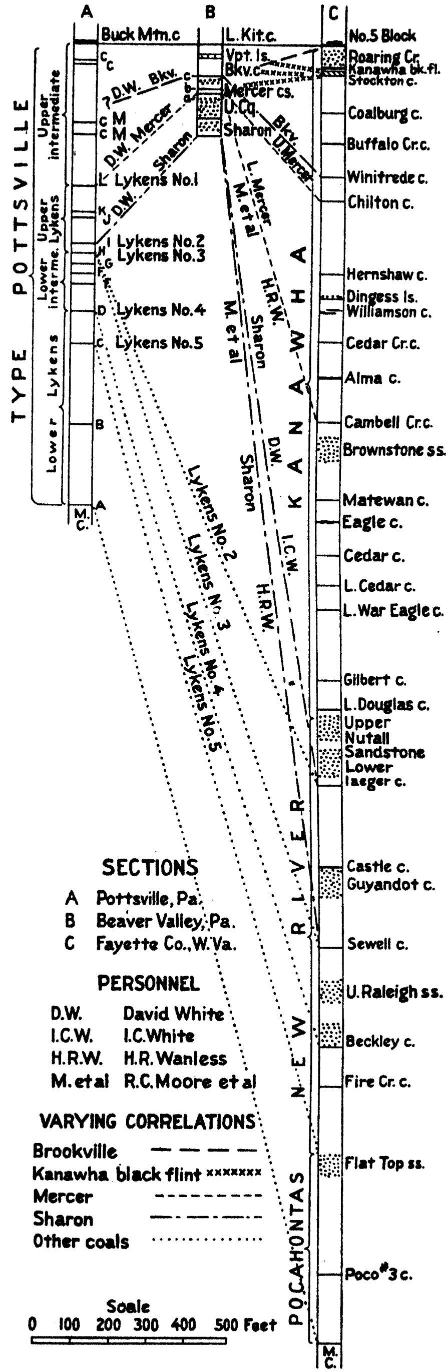


FIG. 5.—Pottsville sections in the Southern Anthracite field of Pennsylvania, Beaver Valley, Pennsylvania, and southern West Virginia, to show the variable thickness and content of the series. The letters *a*, *b*, *c*, show the position of the Brookville (*a*), Upper Mercer (*b*), and Lower Mercer (*c*) coals, all of which have been called "Brookville" in recent literature on western Pennsylvania.

as to the position in the section of what he called the "Brookville coal." In his many references to it the Brookville coal is always above the Tionesta sandstone. The three Porter coals appear to be the Upper Mercer in two benches and the Lower Mercer, or the Tionesta, and Upper and Lower Mercers. Rogers' boundary corresponds to the paleobotanical boundary as drawn by David White.

ROGERS' SECTION, LOWER ALLEGHENY RIVER COAL MEASURES*

FORMATION XIII (LOWER PART)

	Ft.	In.
Kittanning coal	3–4	0
Slate and shale	30	0
Iron ore	4 in. to 6	0
Ferriferous limestone	15	0
Slate and shale	30	0
Clarion coal	3–4	0
Slate and shale	25	0
Brookville coal	1–2	0
Slate and shale	5–15	0
Tionesta sandstone (massive)	50–60	0
Brown and black shale	2–25	0
Tionesta or Mercer coal	1–4	0
Shale and argillaceous limestone	15	0
Upper Porter coal	1	3
Shale and argillaceous sandstone	15	0
Middle Porter coal	1	0
Shale and argillaceous sandstone	15	0
Mercer limestone	2	0
Lower Porter coal	1	8
Shale and argillaceous sandstone	15	0
Seral sandstone and conglomerate (Formation XII)	100	0

* *Ann. Rept.* (1858), Vol. II, pp. 476–77.

SECOND GEOLOGICAL SURVEY

Then in 1874 came the Second Geological Survey of Pennsylvania, under Lesley. In report Q,[7] I. C. White introduced the name "Beaver River Group" to include in the Beaver River Valley the strata from the top of the Homewood (Tionesta) sandstone to the top of the Sharon sandstone, but with this qualification by Lesley: ". . . . provided that the term has the same systematic value as those of the *Freeport*, *Kittanning*, and *Clarion Groups*, and that the group is the *lowest part* of the well-established *Allegheny River Series*.—J. P. L." (italics Lesley's). This would carry the Allegheny River series down to the top of the Sharon sandstone. It was then thought that the Sharon conglomerate might be the top of Formation X or Pocono (No. XI, Mauch Chunk, is missing in that part of the state).

In volume QQ,[8] I. C. White makes no mention of the Allegheny River series, but of the "Lower Productive Coal series," including from the top of the Upper Freeport coal to the top of the Tionesta (Homewood) sandstone, followed by "the Conglomerate Measures, No. XII," identical with the Beaver River group of the preceding volume. Of the Brookville coal, White says (p. 51):

> Below the Clarion, we generally find nothing but sandy shales until we come down to the Tionesta, but sometimes we get a coal resting immediately upon the latter stratum, and I have identified it with the Brookville bed of the First Survey. The bed occurs at an interval of 53 feet below the Ferriferous limestone at this locality, and it is never more than that, but sometimes less.

This agrees with Rogers' position of the Brookville (see above) but does not agree with Rogers' top of Formation XII. It may be explained that at that time I. C. White's work had been confined to the Beaver Valley, where the Homewood sandstone is commonly massive and conspicuous. In the Allegheny Valley that sandstone is generally much less massive, and the conspicuous sandstone is the underlying Connoquenessing sandstone (Formation XII of Rogers),

[7] *Report of Progress in the Beaver Bituminous Coal-Fields of 1878*, p. 66.

[8] *The Geology of Lawrence County, 1879*, p. 27.

as in his section. I. C. White[9] was more impressed by the sandstones than by the accompanying coal beds, which, as a rule, are not impressive.

The confusion existing in the late seventies is shown by W. G. Platt's report H4,[10] which defines the Allegheny River coal series as including the Lower Barren Measures and down to and including the Pocono sandstone (Vespertine). The section includes the "Lower Productive Coal Measures" and "Pottsville conglomerate (Seral), XII." The "Allegheny River series" is not used in the later Q and H reports, which cover the area in question. In those reports "Lower Productive Coal Measures" and "Pottsville—No. 12" were used, the former including strata down to the top of the Pottsville conglomerate, of which the Homewood sandstone was considered to be the top member with the Brookville coal just above. This is distinctly different from Rogers' practice (see above), in which he put the top of Formation XII under the Mercer coals. However, by 1885, the use of the base of the Brookville coal or the top of the underlying Homewood (Tionesta or Piedmont) sandstone had been accepted not only in Pennsylvania but in Ohio as well.[11]

In 1895 the work of the Second Geological Survey of Pennsylvania was summed up in a *Final Summary Report* by Lesley and others, where in Volume III, Part I, page 1920, it says: "The bottom of the Buck Mountain or Red Ash bed, the first coal bed overlying the conglomerate, has been taken as the upper limit of XII." This is definite enough. The section in the bituminous coal fields is given in Volume III, Part II, page 2156. The basal part of the "Lower Productive or Allegheny River series, No. XIII" is as follows:

LOWER PART OF "ALLEGHENY RIVER SERIES, NO. XIII" (D'INVILLIERS)

Kittanning lower coal bed B
Kittanning sandstone
Buhrstone iron ore
Ferriferous limestone
Clarion coal bed A′
Brookville coal bed A
Homewood sandstone: Top of Conglomerate No. XII

This section is strictly in accord with the Rogers section so far as the "Homewood sandstone" is concerned. But it differs in putting the boundary between the Allegheny River series and Formation XII below the Brookville coal rather than below the underlying Mercer coals, as was done by Rogers.

UNITED STATES GEOLOGICAL SURVEY

In 1891 appeared *U.S. Geological Survey Bulletin 65*, "Stratigraphy of the Bituminous Coal Fields of Pennsylvania, Ohio, and West Virginia," by I. C. White, in which Allegheny River series is revived as synonymous with Lower Coal Measures, including down to the Pottsville conglomerate (XII) of which the Homewood sandstone formed the top. Apparently I. C. White assumed that Roger was in error in placing the boundary below the Mercer coals.

In 1904 the U.S. Geological Survey published the Kittanning folio; and in 1906, *Bulletin 279* on the "Economic Geology of the Kittanning and Rural Valley Quadrangles," both by Charles Butts, in which the Pottsville is described as including the Connoquenessing sandstone, Mercer shale, and Home-

[9] "Stratigraphy of the Bituminous Coal Fields of Pennsylvania, Ohio, and West Virginia," *U.S. Geol. Surv. Bull. 65* (1891), p. 19.

[10] *Report of Progress in Indiana County, 1878*, p. 35 (2d Geol. Surv. of Pa.)

[11] Edward Orton, *Ohio Geol. Surv.*, Vol. V: *Economic Geology* (1884), pp. 1–8.

wood sandstone (pp. 30–31). The Allegheny formation contained, from the bottom up: the Brookville coal, Craigsville coal, Clarion sandstone, Clarion clay, Clarion coal, Vanport (Ferriferous) limestone, Kittanning sandstone, Kittanning fire clay, Lower Kittanning coal, and so on (pp. 32–35). This area, studied by Butts, lies immediately southwest of Brookville, and it might be presumed that "Brookville coal' of the Kittanning quadrangle is the true Brookville. Butts described the three lower coals as follows: Brookville coal, "a generally worthless coal lying from 10 to 20 feet above the top of the Pottsville"; Craigsville coal, "two miles northwest of Craigsville is a coal 3 feet thick 50 feet below the Vanport limestone" (it was seen at only two other places in the two quadrangles); Clarion coal, "so far as known, probably rarely exceeds a foot in thickness throughout the quadrangles."

The Kittanning-Brookville area is the type locality for the basal part of the Allegheny formation. In view of the inconsequential character of the three beds just described, is it any wonder that there was confusion as to which bed was "the lowest workable coal seam"?[12] Note that, as against Rogers' final section, in which the "Brookville coal" is 60 feet or less below the Vanport limestone, in the Kittanning, Rural Valley, Foxburg, and Clarion folios, the "Brookville coal" averages from 85 to 112 feet or more below the Vanport limestone, while the Craigsville coal is in the position of Rogers' Brookville coal. It was this difference that first led the writer to question whether the Craigsville coal was not the Brookville coal of Rogers and of the Beaver Valley, where locally it is called the "Pardoe coal." But if the Brookville coal of the type area is one of the Mercers, it means that the "Brookville coal" of Beaver Valley, Southwestern Pennsylvania, Ohio, or northern West Virginia is not Brookville, and sections throughout that area must be revised. Before calling for such a revision, however, let us study the type area more closely.

[12] Rogers, *4th Ann. Rept.*, p. 198.

FOURTH GEOLOGICAL SURVEY OF PENNSYLVANIA (1919——)

The present Geological Survey has taken four steps toward solving this problem.

1. In 1921 B. Coleman Renick[13] traced the strata from the type section in Beaver Valley to Brookville in great detail, leading to the conclusion that

> the sandstone known as the Clarion is in reality the Homewood sandstone of the Beaver River section, and the sandstone called "Homewood" in the Foxburg-Clarion folio is the upper Connoquenessing sandstone of the Beaver River section. The coal called "Craigsville" coal along Allegheny River is the stratigraphic equivalent of the Brookville coal of the Beaver River section.

Because of two weak links in the chain of sections, no change was made in the stratigraphy at that time, pending the strengthening of those links.

2. From 1931 to 1934 Charles K. Graeber[14] surveyed the Brookville quadrangle, with results shown in Table 1. In 1936 and later R. E. Sherrill and others[15] mapped the Hilliards, Franklin, and Oil City quadrangles, tracing, with similar

[13] "The Correlation of the Lower Allegheny-Pottsville Section in Western Pennyslvania," *Jour. Geol.*, Vol. XXXII, No. 1 (1924), pp. 64–80.

[14] "Structure and Stratigraphy of the Brookville Quadrangle," *Pa. Geol. Surv. Bull. 120* (1939), p. 6.

[15] "Geology of the Oil and Gas Fields of the Hilliards Quadrangle," *Pa. Geol. Surv. Bull. 122* (1939), p. 4; "Geology of the Oil and Gas Fields of the Franklin Quadrangle," *Pa. Geol. Surv. Bull. M24* (1941), p. 7; "Geology of the Oil and Gas Fields of the Oil City Quadrangle," *Pa. Geol. Surv. Bull. M25* (1943), p. 7.

TABLE 1

TABLE OF INTERVALS OF LOWER ALLEGHENY GROUP IN PENNSYLVANIA ABOVE OR BELOW THE VANPORT LIMESTONE (IN FEET)
(Lower Kittanning Above; the Others Below)

Source of Sections	Lower Kittanning Coal	First Coal	Second Coal	Third Coal	First Sandstone	Fourth Coal	Fifth Coal	Second Sandstone
1858, Beaver Valley, Rogers	Kit. 45	....	Cl. 33	Bkv. 58	Tion. 68	U.M. 133	L.M. 185	Seral 200
1880, Beaver Valley, L. C. White	Kit. 60	Sg. 5	Cl. 21	Bkv. 53	Hw. 56	U.M. 103	L.M. 132	U.Cq. 157
1905, Beaver Valley, DeWolf	L.K. 56	U.Cl. 5	L.Cl. 25	Bkv. 58	Hw. 60	U.M. 99	L.M. 130	U.Cq. 134
1921, Beaver Valley, Renick	....	U.Cl. 13	L.Cl. 34	Bkv. 57	Hw. 58	U.M. 92	L.M. 119	U.Cq. 128
1921, Slippery Rock, Renick	....	U.Cl. 10	L.Cl. 31	Bkv. 49	Hw. 50	U.M. 81	L.M. 109	....
1939, Hilliards quadrangle, Sherrill	L.K. 45	U.Cl. 20	L.Cl. 35	Bkv. 60	....	....	Mer. 115	....
1911, Foxburg quadrangle, Shaw	L.K. 45	U.Cl. 14	L.Cl. 36	Cr. 64	....	Bkv. 102	....	Hw. 112
1904, Kittanning quadrangle, Butts	L.K. 43	....	Cl. 20	Cr. 50	....	Bkv. 85	....	Hw. 127
1921, Allegheny Valley, Renick	....	U.Cl. 12	L.Cl. 34	Bkv. 71	Hw. 71	U.M. 84	L.M. 117	U.Cq. 130
1910, Clarion quadrangle, Lines	L.K. 32	....	Coal 25	Coal 45	....	Bkv. 98	....	Hw. 108
1939, Brookville quadrangle, Graeber	L.K. 48	....	U.Cl. 32	L.Cl. 45	Cl. 52	Bkv. 84	....	Hw. 88
1944, Brookville quadrangle, Ashley	L.K. 45	....	Cl. 30	Bkv. 47	Hw. 55	U.M. 81	L.M. 115	U.Cq. 125

* U. = Upper; L. = Lower; Bkv. = Brookville; Cl. = Clarion; Cr. = Craigsville; Cq. = Connoquenessing; Hw. = Homewood; K. *or* Kit. = Kittanning; M. = Mercer; Sg. = Scrubgrass; Tion. = Tionesta.

conclusions, the beds on the surface and filling one gap in Renick's work.

3. In 1944 the writer made certain field studies around Brookville. These results can be summarized in Table 1. Note here that the Lower Kittanning coal, with one exception, lies between 45 and 60 feet above the Vanport limestone. Below the Vanport are five coal horizons and two sandstone horizons, each in its own zone. The first coal is from 5 to 20 feet below the Vanport limestone; the second, from 20 to 35 feet; the third, from 45 to 65 feet; the fourth, from 80 to 105 feet; the fifth, from 110 to 135 feet. The matter is presented graphically in Figure 6.

Table 1 and Figure 6 show that the Brookville coal throughout this area was from 45 to 65 feet below the Vanport limestone and that the nomenclature of the Kittanning and later folios in that immediate area was in error.

Renick's work was of a very high order, the valley banks being mapped in great detail for every rock outcrop. His conclusions would have been accepted at that time but for two minor breaks—one over the crest of the Beaver-Allegheny divide, the other between Redbank and Brookville.

Sherrill's work on the coals was one phase of his study of oil and gas sands of Venango County. In a letter of October 2, 1943, Sherrill, after 7 years' work in that area, says:

> I have found no reason in my later work (Oil City) to deviate from my earlier conclusion that the Mercer of Mercer County is the Brookville of the Foxburg-Clarion. Whether or not this is the Brookville at Brookville is a question of which I have no evidence beyond the literature.

WHAT IS BROOKVILLE COAL AT BROOKVILLE?

The Brookville coal was first mentioned in Rogers' final report (1858), in the table already quoted under "First Geological Survey of Pennsylvania." On page 490 of the same report occurs the following:

> The Brookville Coal.—This bed, lying immediately upon, or a few feet above the Tionesta sandstone, has an irregular outcrop, coinciding in the main with that of the Clarion coal and Ferriferous limestone above it, and it will be best described with theirs.

In the same volume (p. 583) the Brookville coal is described as

> seen upon the hills back of the town (Brookville), and on the turnpike towards Clarion. It outcrops towards the fifth axis about 5 miles W. of Brookville, and measures 3 or 3½ feet in thickness; also around the sides of a high hill, 6 miles from Brookville on the Warren Road, 60 or 80 feet below the summit, upon which lie the fragments of the Ferriferous Limestone-bed and and its valuable ore.

FIG. 6.—Sections showing the relations of the sub-Lower Kittanning coals from Mercer County to Clearfield County.

1.* Lawrence Co., Pa., present survey, *A5*, Newcastle quadrangle (24 secs.) (Pls. VI, VII, and VIII).

2. Mercer Co., Pa., *2d Pa. Geol. Surv. Q3* (27 secs.).

3. Northern Butler Co., Pa., *Geol. Surv. Bull. 122*, Hilliards quadrangle (Gen. Sec., p. 4).

4. Clarion Co., *U.S. Geol. Surv. Bull. 454*, Emlenton and Foxburg quadrangles (5 secs. Pls. VI and VIII).

5. Clarion Co., *U.S. Geol. Surv. Bull. 454*, east of Callensburg, Foxburg quadrangle.

6. Clarion Co., southern ½ Clarion quadrangle, Pa., *Pa. Comm. Surv. Rept. 3*, p. 21 (Gen. Sec.).

7. Clarion Co., Limestone Twp., Brookville quadrangle, Pa., *Pa. Geol. Surv. A54*, Pl. 10, Fig. 2.

8. Jefferson Co., Summerville, Brookville quadrangle, Pa., *Pa. Geol. Surv. A54*, p. 47.

9. Jefferson Co., near Conifer, Brookville quadrangle, Pa., *Pa. Geol. Surv. A54*, Pl. 10, Fig. 5.

10. Jefferson Co., southeast of Brookville (Graeber and G. H. A.).

11. Clearfield Co., McGees Mills, Pa., *Pa. Geol. Surv. A65*, Punxsutawney quadrangle, pp. 29 and 30.

12.* Clearfield Co., top ⅓, Clearfield; bottom ⅓, Clearfield Creek, Pa., *Pa. Geol. Surv. Bull. M6*, Part II, p. 158.

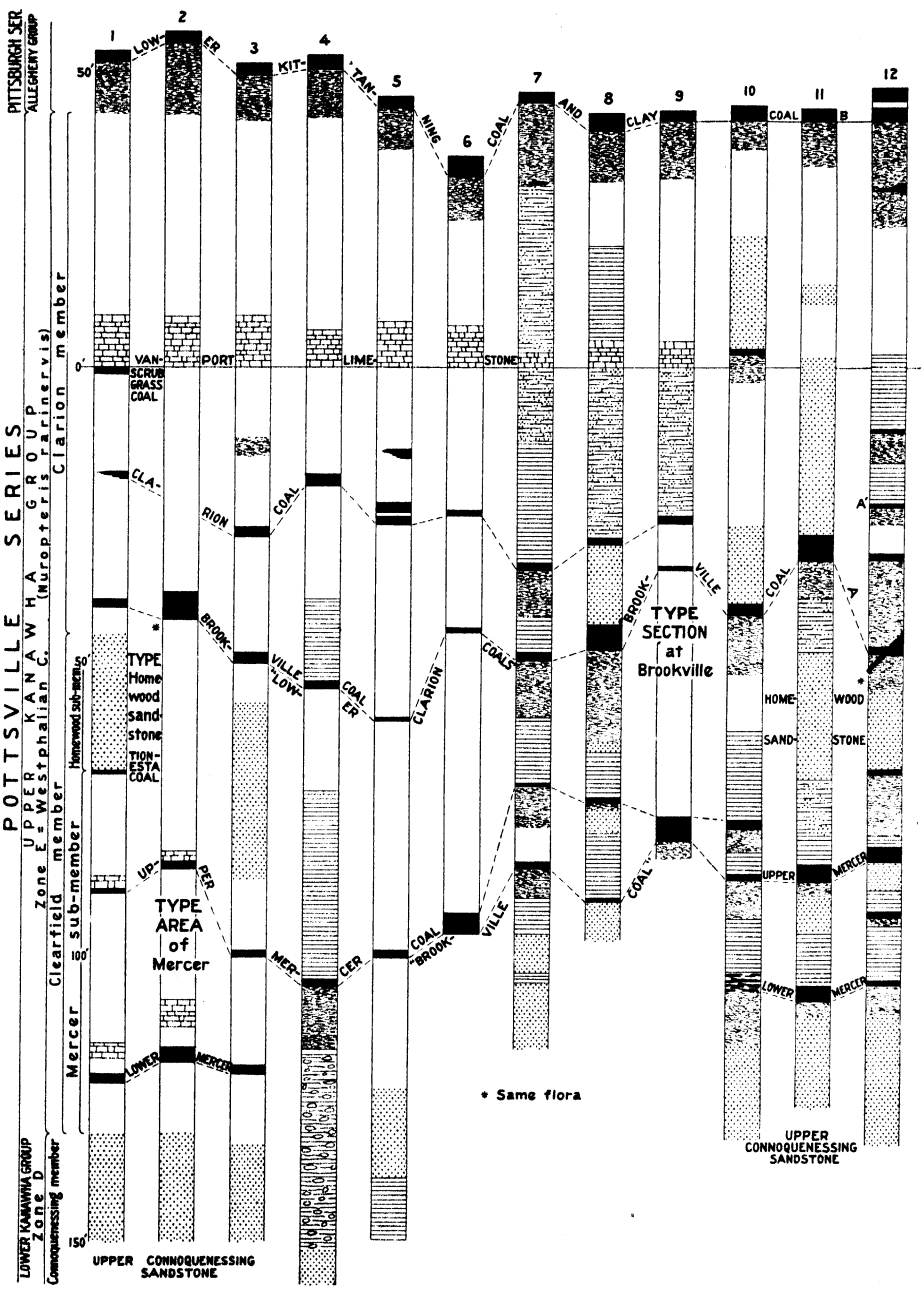


FIG. 6.—See legend on facing page

Again (p. 584): "One mile below Troy (Summerville) the following section was obtained":

	Ft.	In.	Total Feet
Kittanning coal	2	..	2
Brown shale	30	..	32
Ferriferous limestone	4	..	36
Shale and slate	30	..	66
Clarion coal	1	2	67
Shale	55	..	122
Brookville coal	3	6	126
Shale and slate to bed of creek	45	..	171

to the roadside, revealed only about 18 inches of coal. The Clarion seam, 1 foot thick, is exposed at the water trough. The Ferriferous limestone is crossed just beyond that place.

Graeber[17] says:

There is a caved coal bank along the Lakes to Sea Highway in Brookville at the turn of the road near the garage and filling station at an elevation of 1,450 feet. This is probably the coal to which Platt refers and is probably the Lower Clarion coal. The 1-foot coal 30 feet above it is the Upper Clarion. The Vanport limestone has been dug in this hill at about 1,500 feet elevation.

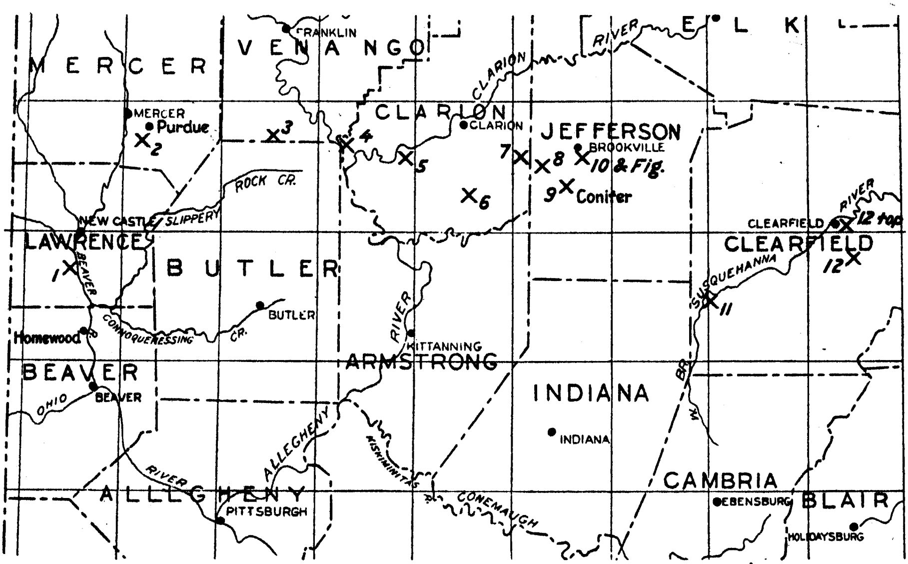


FIG. 7.—Sketch map showing location of sections in Fig. 6

"Seen on the hills back of town" is quite indefinite, as modern mapping shows four beds in the hills around Brookville above the level of its main street. "On the turnpike toward Clarion" is a little more definite. W. E. Platt[16] says:

Proceeding west from Brookville by the Corsica road the Homewood sandstone is first crossed at the outskirts of the town. The Brookville coal seam outcrops at Mr. Braden's house. . . . . An opening once made upon it here, close

This places the Brookville ("Lower Clarion") coal 50 feet below the Vanport limestone.

It is reasonable to assume that the coal described by Platt and Graeber is at the horizon of the coal "seen upon the hills back of town" and is the original Brookville coal at Brookville, 50 feet below the Vanport limestone. The section at Troy obviously lacks the Brookville

[16] "Jefferson County," *2d Geol. Surv. Pa. Rept.* (1881), p. 106.

[17] *Ibid.*, p. 50.

coal; and Rogers' assumption that the coal next below the Clarion was the Brookville, regardless of interval, was erroneous.

Remapping the area about Brookville, based mainly on Graeber's field notes, rechecked by hand level, and new road cuts revealed four coal beds below the level of the Vanport limestone. As the limestone is lacking for a considerable area immediately around Brookville, the usual three-dimension structural pattern was used in which every outcrop of coal must fit into its place with the structure, as letters must fit into a crossword puzzle. As a starter, all available drill records were plotted to get average intervals, as shown in Figure 8.

Intervals obtained from Figure 8 were then supplemented by field measurements. In the end the outcrop of the Lower Kittanning coal must everywhere be 50 feet above the hypothetical positions of the lacking Vanport limestone, and the four other coals at approximately 30, 47, 81, and 115 feet below the hypothetical position of the Vanport. The actual interval used varied from place to place according to available local information, such as drill records or several beds in the same hill. According to this mapping, the 4½-foot coal mined at the Pawneee and Yates mines, 3 miles south of Brookville, at 85 feet below the Lower Kittanning, is the Brookville coal; and the equally thick coal mined at Conifer, and 120 feet below the Lower Kittanning coal, is the Upper Mercer. It is obviously the Conifer (Upper Mercer) coal that Rogers called "Brookville" at Troy (Summerville) and that has been called "Brookville" in the Kittanning, Rural Valley, Foxburg, and Clarion folios and in the *Brookville Atlas*. Changing it in these five areas will be vastly simpler than to accept the "Brookville" of those quadrangles and

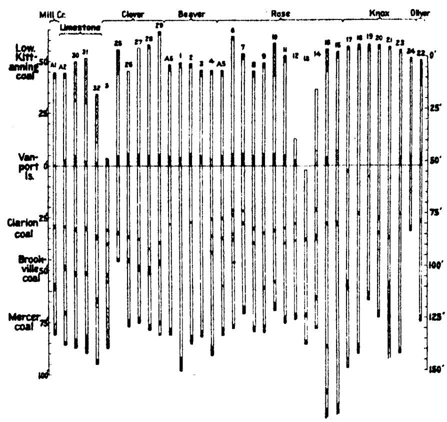


FIG. 8.—Skeleton sections of core-drill records in Brookville quadrangle. Names at the top are of several townships in which the drilling occurs. (Brookville is in Rose Township.) Numbers refer to section numbers in the *Brookville Atlas*.

require the revision of "Brookville" throughout western Pennsylvania, Ohio, northern West Virginia, and Maryland.

If this proposal is accepted, I propose that, until thoroughly established, the Allegheny group, down to the base of the Lower Kittanning coal, be called "Allegheny group, restricted," and that the Pottsville series, including all strata up to the base of the Lower Kittanning coal, be called "Pottsville series, extended."

AGE RELATIONS OF STANLEY AND JACKFORK FORMATIONS OF OKLAHOMA AND ARKANSAS

CHALMER L. COOPER
Illinois State Geological Survey

ABSTRACT

The Stanley and Jackfork formations were recently correlated with the Pennsylvanian but their exact age is still a problem. The flora obtained from these formations is unlike any known flora in North America and is younger than that from the Wedington sandstone member of the Fayetteville shale (Chester) and older than that in Baldwin coal member of the Bloyd shale of the Morrow group (middle Pottsville). White considered the Stanley and Jackfork floras similar to those described from the Ostrau and Waldenburg series of central Europe, which are no younger than lower Namurian (E and possibly H zones).

INTRODUCTION

The Stanley-Jackfork series has been given various age designations since it was first described as Ordovician by J. A. Taff[1] because of limestone boulders of that age in a black-shale formation immediately overlying the Jackfork. The Stanley lies upon the upper member of the Arkansas novaculite, which is considered to be Devonian(?) by the United States Geological Survey, Devonian by some geologists, Devonian and Mississippian by others, and Mississippian by still others. The Jackfork is overlain by the Johns Valley shale or, where this is absent, by the Atoka formation. An accurate determination of the age of the Stanley-Jackfork has not been possible because of the complexity of the structure and the stratigraphy of the Ouachita Mountains and adjacent areas. However, the true Carboniferous age of the series was soon recognized, and for many years the formations were thought to be Mississippian.[2] This belief was influenced by the presence of Mississippian Caney shale[3] in Johns Valley, Oklahoma, overlying the Jackfork. But it is now recognized that the fragments of Mississippian shale at this locality are as truly boulders in the Johns Valley shale as are the large masses of limestone ranging in age from Ordovician to Pennsylvanian. The Johns Valley shale has been correlated with the Morrow group by Bruce H. Harlton,[4] although boulders of this age have been found in the shale. Since the beds in actual contact with the Stanley-Jackfork range in age from middle Pottsville above to Kinderhook (or Devonian?) below, they provide little or no definite evidence on the exact age of the formations.

A Pennsylvanian age for the formations has been suggested by a number of writers, although based on meager and seemingly inconclusive evidence. In 1927 a serious effort was made by Miser and others ('ncluding the writer) to secure sufficient fossil material to place a definite age on the series, and additional

[1] *U.S. Geol. Surv., Geol. Atlas, Atoka Folio 79* (1902), p. 4.

[2] Charles N. Gould, "Index to the Stratigraphy of Oklahoma," *Okla. Geol. Surv. Bull. 35* (1925); Hugh D. Miser and Charles W. Honess, "Age Relations of the Carboniferous Rocks of the Ouachita Mountains of Oklahoma and Arkansas," *Okla. Geol. Surv. Bull. 44* (1927).

[3] George H. Girty, "The Fauna of the Caney Shale of Oklahoma," *U.S. Geol. Surv. Bull. 377* (1909).

[4] "Micropaleontology of the Pennsylvania Johns Valley Shale of the Ouachita Mountains, Oklahoma, and Its Relationship to the Mississippian Shales," *Jour. Paleon.*, Vol. VII (1933), pp. 3–29.

collections were made during the years 1928–31. These collections, together with those obtained during previous years, were studied by David White (plants), Girty (invertebrates), and Harlton (microfossils). All of the available data have been ably presented by Miser.[5]

PALEOBOTANICAL EVIDENCE

The most useful of these collections yet obtained for correlative purposes are the floras. However, it should be borne in mind that systemic boundaries are drawn at different stratigraphic horizons when based on different classes of fossils. O. H. Schindewolf,[6] for example, pointed out that the Devonian-Carboniferous boundary, drawn on flora evidence, was placed in middle Tournaisian (about the base of the Osage), while, on the basis of corals, it was placed below the top of the Famennian, between Oberdevon Stufe IV and V. The line, based on brachiopods and trilobites, was placed between these two extremes, the latter approximating the base of the Kinderhook.

The Stanley-Jackfork floras formed the basis of two papers by David White,[7] the last of which was published, practically without modification, after his death early in 1935. In this paper the floras of the Stanley-Jackfork were compared with those from the Morrow, with the Wedington and other Chester floras, all from this country, and with the Upper Carboniferous floras of Upper and Lower Silesia, the Ostrau and Waldenburg series.

The Stanley-Jackfork, in spite of careful collecting for more than a quarter-century, had yielded only 34 species of plants—26 from the Stanley and 16 from the Jackfork (8 being common to the two formations). Of the 34 species, 19 are new and 8 are listed without specific designation. Regardless of the number of new species, no less than 18 from the list were closely compared to European species from the lower Upper Carboniferous, and the fauna was regarded as sufficiently comprehensive to determine the Pennsylvanian or Mississippian age of the Stanley-Jackfork with reasonable certainty.

The most extensive known Chester floras were obtained from the Wedington sandstone member of the Fayetteville shale of Arkansas,[8] the Chester series of the Mississippi Valley, and (unpublished) from the Appalachian section in West Virginia, Virginia, and Pennsylvania. D. White[9] considered the Wedington flora to consist of forms identical with or closely allied to others characteristic of the upper Chester; of unique forms; and of plants related to Pennsylvanian forms. He found two species, *Rhodea goepperti* and *Sphenopteris mississippiana*, common to the Wedington and Stanley-Jackfork floras; a lepidophyte, *Lepidodendron* cf. *L. wedingtonense*, was provisionally placed with this Wedington species. *Sphenopteris mississippiana* is probably the same as *S. hoeninghausii* of the Namurian of

[5] "Carboniferous Rocks of Ouachita Mountains," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XVIII (1934), pp. 981–85.

[6] "Die Liegendgränze des Karbons im Lichte biostratigraphischer Kritik," *Compt. rend. Cong. advance. des ét. de stratigr. carbonifère, Heerlen, 1927* (Liége, 1928), pp. 651–61.

[7] "Age of Jackfork and Stanley Formations of Ouachita Geosyncline, Arkansas and Oklahoma, as Indicated by Plants," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XVIII (1934), pp. 1010–17.

[8] White, "Fossil Flora of the Wedington Sandstone Member of the Fayetteville Shale," *U.S. Geol. Surv. Prof. Paper 186* (1937), p. 51.

[9] "Fossil Plants from the Stanley Shale and Jackfork Sandstone in Southeastern Oklahoma and Western Arkansas," *U.S. Geol. Surv. Prof. Paper 186* (1937), pp. 43–52.

Europe. As a result of his comparison of the Stanley-Jackfork floras with known Chester floras, White concluded that there is "little or no doubt that the Stanley as well as the Jackfork is younger than the entire Fayetteville shale and its Mississippian correlatives."

One of the best-known Carboniferous floras in North America is found in the roof of the Baldwin coal member of the Bloyd shale of the Morrow group in Washington County, northwestern Arkansas. This flora, first studied by Lesquereux in 1884, comprises about 125 species, which can be correlated with such middle Pottsville formations as the shales above the Lookout sandstone in Alabama and Tennessee, the Norton of Virginia, the Sewell of West Virginia, and the Caseyville of Illinois and Kentucky. A comparison of this flora with that from the Stanley-Jackfork shows that "the deposits are distinctly older than the plant-bearing shale of the Morrow group, including but two species that have been definitely recognized as present also in the shale." The Jackfork is then "referred to the early Pottsville, earlier than the coal-bearing shale of the Morrow group—specifically the lower Pottsville. . . . . Probably the whole of the Stanley shale falls in the same unit of the lower Pennsylvanian," regardless of the fact that "little is known either as to the flora of that portion of the Morrow group lying below the coal-bearing shale or as to marine faunas in the lower Pottsville of the eastern basins." Therefore, the Stanley-Jackfork flora, by comparison with known floras in formations not in actual contact with these formations, is "distinctly older than the plant-bearing shale of the Morrow" and "younger than the Wedington sandstone member" of the Fayetteville formation—in other words, older than middle Pottsville and younger than middle Chester.

Earlier, White was of the opinion that the formations were Mississippian,

> based largely on the parallelism in floral composition and the presence of closely related or identical floral elements in the Jackfork and Stanley and in the fully elaborated floras of the Waldenburg and Ostrau series (Culm) of the former Austro-Silesian region, which were then regarded as uppermost Lower Carboniferous.

Regardless of the position of these beds with reference to the European Upper and Lower Carboniferous boundary, the important fact is that the Stanley-Jackfork flora was considered to be most closely correlative with a European assemblage and not closely comparable with any known flora from either Mississippian or Pennsylvanian formations in America.

It should be pointed out that White's designation of the Pennsylvanian age of the Stanley-Jackfork was influenced by the transfer of certain beds of the Lower Carboniferous of Europe into the newly created Namurian series, the lower division of the Upper Carboniferous, by action of the Congrès pour l'avancement des études de stratigraphie carbonifère, which met at Heerlen, Netherlands, in 1927. He also "accepted the thesis of the essential equivalence of the American Mississippian and Pennsylvanian, respectively, to the European Lower and Upper Carboniferous, as revised by the Heerlen Congress of 1927.—EDITOR."[10]

INVERTEBRATE EVIDENCE

Invertebrates are extremely rare in the Stanley and Jackfork. A small collection made by Honess was regarded by Schuchert and Ulrich (each of whom examined the fossils independently) as

[10] *Ibid.*, p. 46.

inconclusive; but each favored a post-Chester age, possibly early Pennsylvanian.

The largest collection of invertebrates was by Miser[11] from the Jackfork near Little Rock and identified by Girty. Of the 28 forms listed by Girty from this fauna, 16 were given only generic designation (three-fourths of these were queried); and all of those given specific designation were qualified either by "?" or "aff." If the species themselves (or even genera) cannot be definitely recognized, it follows that their reference to known species described from other formations (either Mississippian or Pennsylvanian) must be regarded as uncertain.

Harlton[12] lists but does not figure a few species of microfossils from the Stanley-Jackfork. One, *Healdia caneyensis*, is known from the Chester of Illinois, and the similarity of other genera to Chester genera was recognized by him.[13] The Pennsylvanian aspect of most all late Chester faunas is well known.

All who have tried to solve the perplexities involved in the correlation of borderline formations are well aware of the great weight of evidence necessary for conclusive age determinations due to the early inauguration of new species and to the holdover of old species. For these reasons the invertebrate evidence produced so far from the Stanley and Jackfork should be regarded as inconclusive. A promising field for search for this evidence is in the upper Mauch Chunk beds of West Virginia and Virginia.

Since the first Heerlen Congress and since the papers of White, Miser, and others appeared, a number of factors have made a reopening of the question of the age of the Stanley and Jackfork seem advisable. The second Heerlen Congress, also devoted exclusively to problems of the Carboniferous, met in 1935, just subsequent to the International Botanical Congress at Amsterdam. Unlike the first congress, at which no American geologist was present, the second was attended by a number of geologists from this continent, among whom were C. A. Arnold and R. C. Moore. Much work had been done in the intervening eight years, as evidenced by the large number of papers in the *Compte rendu*.

At the first session in 1927 many delegates were of the opinion, as shown by their correlation charts, that the Ostrau and Waldenburg series occupied all of the Namurian (and probably more).

> Thus in the Namurian, defined as it has been agreed, it would be necessary to include the Waldenberg beds or strata and certainly a part of the Ostrau beds or strata. It is not impossible that the upper Ostrau beds belong to the Westphalian. Some delegates consider this conclusion even probable. As for the definite solution, it will result from new researches based on more complete materials and especially more systematic collections, materials which, actually, are totally missing in that which concerns the Ostrau beds.[14]

One delegate, Arnold Markowski,[15] even went so far as to include the Dinantian with our Pennsylvanian! However, no serious attempt was made to correlate the American section with the European section.

The second congress, owing to the

[11] *Ibid.*, pp. 989–91.

[12] "Carboniferous Stratigraphy of the Ouachitas with Special Study of the Bendian," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XVIII (1934), pp. 1035–37.

[13] *Ibid.*, p. 1021.

[14] K. Pattiesky, "Discussion générale," *Compt. rend., Cong. avance. des ét. de stratigr. carbonifère, Heerlen, 1927* (Liége, 1928), p. xxxix.

[15] "Coup d'œil sur la structure du bassin houiller polonais," *Compt. rend., Cong. avance. des ét. de stratigr. carbonifère, Heerlen, 1927* (Heerlen, 1928), pp. 481–85.

presence of well-known American geologists and to the visit of W. J. Jongmans to many important American Carboniferous localities before the sessions, reached a much better understanding of the relationship between the Carboniferous of the two continents. The research on faunas, as well as on floras, had been greatly advanced since the first meeting, as shown by the work on the goniatites, corals, brachiopods, etc.

W. J. Jongmans and W. Gothan[16] in their "considerations of the results" of the second congress, presented a correlation table, which is given in part in Table 1. This table was drawn largely from a paper presented at the congress by R. C. Moore.[17]

TABLE 1

	Europe		North America
STEPHANIAN			Marathon orogeny
	Upper Ottweiler schichten		VIRGIL SERIES
	Middle Ottweiler schichten (red beds, etc.)		
			Arbuckle orogeny
	Lower Ottweiler schichten		MISSOURI SERIES
	Asturian orogeny		*Disconformity*
WESTPHALIAN	D		
	C (Staffordian of England)		DES MOINES SERIES
	B		
	A		
			Wichita orogeny
NAMURIAN	Lower Magerkohle	G_1	
			MORROW SERIES
	Flözleeres	R	
	Erzgebirgian orogeny		*Disconformity*
		H	
			CHESTER SERIES
	Lower Namurian	E_1^2	
DINANTIAN		P_2	
	Disconformity		*Disconformity*

Since, by definition, the base of the European Upper Carboniferous is the base of the Namurian, it is readily seen that this boundary is well below the top of the American Mississippian, somewhere in the Chester series. It should also

[16] "Betrachtungen über die Ergebnisse des zweiten Kongresses für Karbonstratigraphie," *Compt. rend., Deuxième cong. avanc. des ét. de stratigr. carbonifère, Heerlen, 1935* (Maestricht, 1937), p. 22.

[17] "Comparison of the Carboniferous and Early Permian Rocks of North America and Europe," *Compt. rend., Deuxième cong. avanc. des ét. de stratigr. carbonifère, Heerlen, 1935* (Maestricht, 1938), p. 671.

be noted that the E and H zones (*Eumorphoceras* and *Homoceras*) are shown to be Mississippian, corresponding to our upper Chester. Pattiesky[18] stated in his report at the second congress:

> To the E-zone belong the non-bedded 400 m thick Hultschin formation as well as the greater part of the 3,000 m thick Ostrau formation. Only the uppermost horizons of the Ostrau formation are likely to be connected with the H-zone.

His correlation table is reproduced here, in part, as Table 2.

TABLE 2

Age		Formation	Zone
WEST-PHALIAN	V_β	Karwin beds	Doubrava
NAMURIAN	V_α–IV_ϵ IV_α–IV_δ		Suchá Sattelflöz
	IV_1–H	Upper Ostrau beds	Poruba Jakewitz
	IV_1–E to IV_1–S	Lower Ostrau beds	Hrusŏv Petřkovice
		Hultschin beds Upper Wagstadt beds	
VISÉAN	III_γ	Lower Wagstadt beds Gratz greywacke	
	III_β	Posidonien shale (upper part)	

Other European workers substantiate this position of the Ostrau beds, one even including all of this series within the E zone, placing the lower Sattelflöz in the H zone.[19] Eberhard Wirth,[20] after a careful faunal study of the Ostrau, delimited the series by the Sudenten disconformity at the base and placed the upper limit at a "?Discordanz" at the line separating the horizon of *Cravenoceras edalense* from that of *Eumorphoceras bisulcatum*, in the lower part of E_2. This line appears to be in the lower part of the lower Sabdenian of England and within the Elvira group of the type area for the Mississippian in North America. Therefore both Knopp and Wirth place the Ostrau lower in the Namurian than Pattiesky. The Waldenburg series of Lower Silesia represents only the lower part of the Ostrau, according to most reports.

The Chester-Morrow interval has been regarded as of comparatively short duration by most workers. In the type Chester areas around the southern margin

[18] "Das Verhältnis der Zonen von *Diplotema adiantoides* und der Lygnopteris Arten zu den Goniatiten-zonen des ostsüdetischen Karbons," *Compt. rend., Deuxième cong. avance. des ét. de stratigr. carbonifère, Heerlen, 1935*, pp. 719–20.

[19] L. Knopp, "Einige neue Goniatitenfunde im Oberschlesien," *Geol. Vereinigung Oberschlesiens, Jahresber., 1934* (1935), Pl. 2.

[20] "Die faunistische Altersbestimmung der Ostrau Schichten," *Neues Jahrb.*, Beilage-Band LXXIII, Abt. B (1915), Table 7.

of the Eastern Interior basin in Illinois and Kentucky, the oldest Pennsylvanian (probably as old as Morrow) rests on variable thicknesses of the Kinkaid limestone, but locally it overlaps Chester formations as low as Menard. On the east and southeast side of the basin in Kentucky and Indiana a number of channels, some 300 feet deep, have been cut into the Chester by pre-Pennsylvanian erosion.[21] At other places in the central United States the Pennsylvanian rests on successively older formations, down to the Ordovician St. Peter sandstone, a horizon stratigraphically about 5,000 feet below the base of the Pennsylvanian of southern Illinois. While this erosion was going on, deposition occurred in some parts of the continent. It is in these areas that the conclusive evidence of the age of the Stanley and Jackfork and other formations of uncertain affinities, such as the Parkwood of Alabama and the Tensus of Texas, will be found.

The greatest known thickness of Upper Mississippian rocks outside of the type Chester area is found in the Mauch Chunk and upper Greenbrier of southeastern West Virginia and adjoining parts of Virginia. In this area D. B. Reger[22] described an aggregate of almost 4,000 feet of Chester beds in the two series. Even if the Princeton and the Bluestone, which may eventually be shown to be a part of the Pocahontas group, are excluded, nearly 3,000 feet of beds remain in the Upper Mississippian. Recent ostracode studies have shown the fauna of the Reynolds member of the Bluefield formation to be about Menard in age.[23] If the Droop is correlated with the Palestine, there remain about 2,000 feet of beds which cannot reasonably be crowded into the Clore-Kinkaid interval of the type section. When the lower parts of the Chester–Mauch Chunk correlations are taken into consideration, the probabilities are that the several hundreds of feet of upper Mauch Chunk were deposited after Kinkaid deposition or during the post-Chester–pre-Pottsville interval, or the interval between the Pitkin and the Hale of Arkansas.

The middle Pottsville, according to Charles Butts,[24] is equivalent to the New River group of West Virginia and the upper Lee and lower Norton of Virginia. There are approximately 1,000 feet of beds in the lower Pottsville or Pocahontas group. The upper Chester, above the Wedington sandstone in Arkansas, contains about 700 feet of strata. If present correlations with southeastern West Virginia are correct, some 2,000 or more feet of strata lie above the middle Chester in the upper part of the Mauch Chunk series. Here, then, are some 3,000 feet of strata of post-middle Chester and pre-middle Pottsville age, which have no correlatives in the Pitkin-Hale unconformity and in which interval lie the Stanley and Jackfork formations.

CONCLUSIONS

The Pennsylvanian age designation of the Stanley shale and the Jackfork sandstone, on the basis of correlation with the Ostrau and Waldenburg series, classed as European Upper Carboniferous, and upon inadequate invertebrate data, deserves reconsideration because

[21] J. Marvin Weller and A. H. Sutton, "Mississippian Border of Eastern Interior Basin," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXIV (1940), p. 848.

[22] David B. Reger, *W.Va. Geol. Surv. Rept., Mercer, Monroe, and Summers Counties* (1926), pp. 291–491.

[23] Chalmer L. Cooper, "Chester Ostracods in Illinois," *Ill. Geol. Surv. Rept. Inv.* 77 (1941), p. 13.

[24] "Geology of the Appalachian Valley in Virginia," *Va. Geol. Surv. Bull.* 52 (1940), p. 417.

of the information presented by European workers at the second Heerlen Congress and in other publications.

The Stanley and Jackfork formations deposited in a post-Chester–pre-Pottsville interval, although probably younger than any known Chester in the Mississippi Valley type area, are not necessarily Pennsylvanian in age. Whether these formations are Pennsylvanian, equivalent to some part of the 1,000 feet of Pocahontas lying below the Morrow horizon, or are Mississippian, correlative with some part of the 2,000 feet of beds younger than the type Chester, must await the presentation of more pertinent evidence from both the Appalachian and the Midcontinent areas. If correlative with the European E and H zones, which are probably equivalent to our Mississippian upper Chester and post-Chester formations, the Stanley and Jackfork should be considered Mississippian in age.

Considered on a diastrophic basis, the later stages of a period in a region of orogenic disturbance begin with preliminary pulsations which are reflected in the changing character of the formations toward the top of the system. The sediments become more clastic, finally resulting in coarse sandstone with beds of conglomerate. Such was the history of the closing stages of the Ordovician when, after early shale deposition, pulsations of the Taconic orogeny gave rise to the Bald Eagle conglomerate and the Juniata sandstone before the recognized close of the period.[25] Likewise the close of the Devonian was marked by the Acadian orogeny producing the thick Catskill clastics before its culmination. Somewhat similar conditions characterized the Laramide orogeny closing the Cretaceous period. The Stanley is predominantly shale and marks the beginning of the Wichita orogeny, while the Jackfork is predominantly sandstone, marking a later phase of the orogeny. The Morrow and Atoka, which are truly Pennsylvanian, were deposited in the earliest Pennsylvanian sea after the culmination of the Wichita orogeny.

[25] R. T. Chamberlin, "Pleistocene Coast Range Orogenesis in Southern California," *Jour. Geol.*, Vol. LII (1944), p. 358.

AN OCCURRENCE OF GRANITE IN POPE COUNTY, ILLINOIS[1]

J. MARVIN WELLER AND ROBERT M. GROGAN
Illinois State Geological Survey, Urbana

ABSTRACT

Fragments of weathered granite have been found in eastern Pope County, Illinois, in an area of widespread basic igneous intrusives. The origin of the fragments is a matter for conjecture. No parent granitic intrusive has been discovered, but neither has conclusive evidence of glacial origin been found.

INTRODUCTION

The first discovery of an igneous intrusion in Illinois was by W. S. Tangier Smith in 1902. This was the Mix dike of Pope County in the Ohio River bluff, 2 miles north of Golconda. During the course of subsequent field work in 1903 by members of the United States Geological Survey, igneous rock was found in outcrop at three other places, and two more localities were discovered where loose fragments of igneous material occurred. Specimens studied by Albert Johannsen were identified as mica-peridotites and lamprophyres.[2]

The presence of dikes of basic igneous rock intruding coal No. 5 in near-by Saline County was reported in 1907,[3] and more detailed descriptions of these dikes were published in 1919.[4] A sample from one of the mines at Eldorado was identified as olivine kersantite by Dr. Johannsen.

Field work in 1917–19 in preparation for the report on the geology of Hardin County[5] included a restudy of the known igneous occurrences, the discovery of the Sparks Hill intrusion, and the recording of another locality, previously found by U. S. Grant. L. W. Currier studied specimens of these rocks and, in addition to the types identified by Johannsen from that area, recognized the occurrence of "volcanic breccia(?)."[6]

Test borings for oil in 1940, some 3 miles southwest of Omaha, Gallatin County, encountered numerous dikes and possibly sills of mica-peridotite.[7] Outcrops of basic dikes are reported in the southeastern corner of Franklin County.[8]

Field studies beginning in 1941 and connected with wartime fluor spar investigations have resulted in the discovery, on the flanks of Hicks dome, of two new igneous-rock outcrops by R. M. Grogan. New developments by the Aluminum Ore Company have brought to light igneous rock recognized by A. H. Sutton in mine workings and diamond-drill cores in and near the Blue Diggings vein. "Mica dirt" is reported

[1] Published by permission of the chief of the Illinois State Geological Survey.

[2] H. F. Bain, "The Fluorspar Deposits of Southern Illinois," *U.S. Geol. Surv. Bull:* *255* (1905), pp. 27–30.

[3] Frank W. De Wolf, "Coal Investigation in the Saline-Gallatin Field, Illinois, and Adjoining Area," *U.S. Geol. Surv. Bull. 316* (1907), p. 122 (reprinted in *Ill. State Geol. Surv. Bull. 8* [1907], p. 216); "Coal Investigations in Saline and Williamson Counties, Illinois," *Ill. State Geol. Surv. Bull. 8* (1907), p. 241.

[4] G. H. Cady, "Coal Resources of District V," *Ill. State Geol. Surv. Co-op. Min. Ser. Bull. 19* (1919), pp. 34–35, 56–61.

[5] Stuart Weller *et al.*, "The Geology of Hardin County," *Ill. State Geol. Surv. Bull. 41* (1920).

[6] *Ibid.*, p. 237.

[7] R. M. English and R. M. Grogan, "The Omaha Pool and Peridotite Intrusives, Gallatin County, Illinois" (in manuscript).

[8] Carl B. Anderson, Gulf Refining Company, personal communication, October, 1943.

to have been found in auger prospecting along the lower part of Grand Pierre Creek. Finally, weathered granitic fragments were observed by J. M. Weller about a mile southwest of the old Empire mine.

This last occurrence is of particular interest because it may indicate the presence of an igneous intrusion of quite different character from any other known in southern Illinois or the adjacent part of Kentucky.

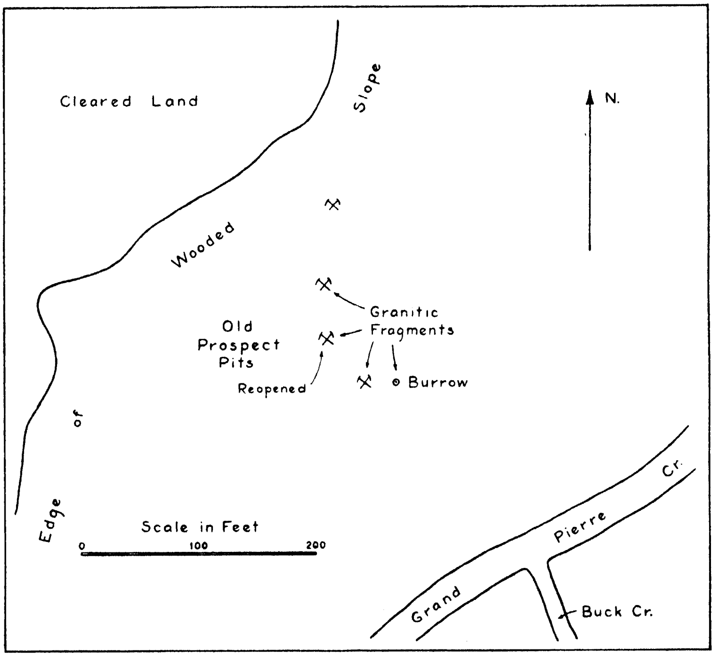


FIG. 1.—Sketch map showing location of granite in NE. ¼, SE. ¼, Sec. 33, T. 11 S., R. 7 E., Pope County.

POPE COUNTY GRANITE

Fragments of much-weathered, light-colored granitic rock and vein quartz containing flakes of muscovite were first seen on the dump of an old prospect pit (Fig. 1) located at an elevation of about 410 feet in the western part of the NE. ¼, SE. ¼, Sec. 33, T. 11 S., R. 7 E., Pope County. The prospect pit is halfway up the slope on the northwest side of Big Grand Pierre Creek nearly opposite the mouth of Buck Creek. Golconda shale and limestone underlie the slope, Cypress sandstone outcrops in the creek bed near by to the northeast, and Hardinsburg sandstone occurs up the hillside to the west.

The pit was later reopened to a depth of about 6 feet through the courtesy of Mr. Ira Adams, of Karber and Adams, at that time operators of the near-by Douglas fluor spar property. It exposed only talus material; but, in addition to the sandstone blocks and fragments of shale and weathered and silicified limestone, it also yielded fragments of granitic rock. The largest piece encountered

was about 1 foot in diameter. A search of the hillside produced several small fragments of similar rock, derived from other prospect pits, both higher and lower on the slope, and from a woodchuck burrow at the base of the hillside.

PETROGRAPHIC CHARACTER

R. M. Grogan has made a preliminary examination of this rock and finds that the only minerals recognizable in hand specimen are quartz and muscovite mica. The granules of quartz are generally less than 5 mm. in length, but some specimens up to 6 cm. long are essentially masses of quartz. The mica crystals are all small, less than 3 mm. across. In most hand specimens a rude linear flow structure is evident.

Microscopic examination of four thin sections shows that the rock consists mainly of quartz, biotite, muscovite, and masses of sericite probably derived from feldspar. Magnetite and secondary hematite form less than 1 per cent of the rock. Apatite, zircon, and tourmaline are rare accessories. No feldspar was observed, although the general nature of the rock and the abundance of sericite suggest that some may have been present originally. There is no definite indication of ferromagnesian minerals other than biotite.

Quartz makes up an estimated 55 per cent of the rock, biotite 10 per cent, muscovite 5 per cent, and sericite most of the remaining 30 per cent.

The quartz is in crystals mostly less than 1 mm. in size, containing numerous lines of tiny inclusions and showing prominent undulatory extinction. Commonly a number of these smaller crystals are interlocked with one another, producing the larger masses visible in hand specimens. The biotite is in laths ranging up to several millimeters in length and is so strongly bleached that only a faint pleochroism can be seen. The muscovite is in laths of similar size but of perfectly fresh appearance.

Most of the sericite is fibrous, many of the fibers being arranged in small fan-shaped aggregates; but occasional larger patches of sericite exhibit more or less uniform extinction and resemble imperfectly formed crystals of muscovite. The sericite areas lie between and around the quartz and biotite in places where primary feldspar would be expected.

Flow structure is more clearly evident in thin section than in hand specimens and more nearly resembles that of the "primary" gneisses than of the truly metamorphic gneisses. Biotite laths, linear groups of quartz crystals, and linear sericitic areas all have their long axes arranged in subparallel fashion. Additional effects of stress appear to be shown by undulatory extinction in the quartz grains and by slight bending and "stretching" of biotite crystals.

The common linearity is not shared by the muscovite, however, whose laths, as often as not, lie at strongly divergent angles. This fact suggests that the muscovite and sericite were formed, probably by hydrothermal action, after the crystallization of the quartz, biotite, and possible feldspar and after the development of the parallel texture.

The granite fragments might be weathered debris from a small dike or plug beneath the site where the fragments were found. Many such small intrusives exhibit flow structure of the type described above. The considerable range in grain size of the quartz is less characteristic of normal small intrusives but might be the result of the hydrothermal activity responsible for the unoriented muscovite and the sericite.

Additional evidence bearing on the

origin of the granite consists of the occurrence near the prospect pits of angular blocks of highly quartzitic sandstone. Some quartzitization of sandstone is commonly present along the faults in Pope and Hardin counties. No fault is known, however, close to this locality; and, moreover, the quartzitization of some of these blocks is more complete than is generally found along the known faults of the region. This suggests that some process other than faulting, perhaps contact metamorphism, may have produced the quartzite.

POSSIBLE GLACIAL ORIGIN

Because no actual exposure was discovered, the existence of a granitic intrusion at this place has not been established, and the alternative possibility that this material is of glacial derivation must be carefully considered.

A granitic boulder in Pope County might have reached its present location by ice-rafting during later Pleistocene time, when the lower Ohio River and its tributary valleys were ponded. Boulders, believed to have been transported in this way, were found by Stuart Weller near Carrsville, Kentucky, across the Ohio River from Rosiclare, Illinois.[9] The Pope County granite, however, occurs 7 miles north of the Ohio River and up to an elevation of about 430 feet. This is 100 feet higher than the present Ohio River, considerably higher than any of the late-Pleistocene terraces present in the region and higher than any ponded stage of the river that is known. The highest terrace recognizable near the mouth of Grand Pierre Creek has an elevation of between 340 and 360 feet. An ice-rafted boulder might have been carried to this position by a sudden rise in level of the Ohio, such as might be occasioned by the release of water ponded behind an ice-jam upstream. The Pope County granite, however, is deeply weathered, whereas known ice-rafted boulders consist of practically fresh rock or have weathered "rinds," ½ inch thick at most.

Granite in Pope County might be part of an outwash deposit derived from a glacial ice lobe to the north. Grand Pierre Creek, however, heads several miles south of the margin of the Illinoian drift sheet, and neither it nor any nearby valley is known to carry outwash. Moreover, the granite occurs upon a hillside some 75 feet above the valley bottom and is more deeply weathered than would be expected of Illinoian outwash.

The most reasonable alternative to a local intrusion as the source of the Pope County granite appears to be a hitherto unrecognized remnant of ancient glacial drift located considerably south of the margin of the Illinoian sheet. A few such remnants have been recognized in southeastern Missouri, southern Illinois, and northern Kentucky.[10] These remnants are mostly located high topographically, in situations where they have escaped erosion. The Pope County granite occurs as fragments on the steep southeastern slope of a knob separated by a low saddle from, and about 20 feet lower than, the prominent ridge west of Grand Pierre Creek. Conceivably, this knob could bear a small remnant of ancient

[9] Frank Leverett, "The Pleistocene of Northern Kentucky," *Ky. Geol. Surv.*, ser. 6, Vol. XXXI (1929), pp. 37 and 43.

[10] Leverett, "Oldest (Nebraskan?) Drift in Western Illinois and Southeastern Missouri in Relation to 'Lafayette Gravel' and Drainage Development" (abstract), *Geol. Soc. Amer.*, Vol. XXXV (1924), p. 69; "Pleistocene Features in Ste. Genevieve County, Missouri," footnote in Geology of Ste. Genevieve County, Missouri, by Stuart Weller and Stuart St. Clair, *Mo. Bur. Geol. and Mines*, ser. 2, Vol. XXII (1928), p. 252; and pp. 33–35 of ftn. 9.

drift concealed by a thin cover of loess, and the weathered nature of the rock would be consistent with such an occurrence. However, a careful search of the dumps adjacent to the old prospect pits, the near-by hillside, and fresh gullies that drain away from the knob in the opposite direction has failed to reveal any other rock fragments that could not have been derived from formations naturally outcropping at this place. If a glacial remnant existed here, it would almost certainly include other erratic material in addition to granite, and other erratics should be found.

CONCLUSIONS

The origin of the Pope County granite has not been established. It cannot be definitely assigned to a local intrusive until the solid granitic parent-mass is exposed. On the other hand, its derivation from a near-by remnant of ancient glacial till is not substantiated by the finding of other associated erratic material.

FURTHER OBSERVATIONS ON THE ORIGIN OF BEACH CUSPS

O. F. EVANS
University of Oklahoma, Norman, Oklahoma

Since the publication of my article, "The Classification and Origin of Beach Cusps,"[1] I have examined scores of miles of shoreline. Although nothing has been found to change the conclusions as given in the first paper, some new relations have been discovered that seem worth recording.

Cusps of Class 1[2] were described as "very large capelike cusps formed during storms along beaches that are susceptible to considerable erosion and deposition." Such storm cusps result from the concentrated work of waves and currents, which causes rapid local erosion at the shoreline, resulting in overloading of the transporting agents, which are then forced to drop a part of their loads a short distance to leeward. The obstruction thus formed turns the shore current outward, and a cuspate form is built.

Such a cusp may remain isolated and unrelated to other cusps, or the waves may refract around its end and cause erosion to its leeward. Thus a second indentation is formed, the currents and waves are again overloaded, and another projection results.

Storm cusps, when first formed, have a short underwater extension at their outer ends, but as current action and beach drifting continue, a subaqueous ridge is sometimes built out for a long distance. This ridge may later be moved inshore by the waves of translation until its material again forms a part of the beach ridge.

Where there is a large amount of sand and the water is shallow with a gentle offshore slope, these cusps with their underwater extensions are nearly perpendicular to the shoreline, and the submerged ridges remain straight. Such shallow-water cusps constitute Class 2[3] and are described as "large cusps which have their apexes continuing out into the lake as a ridge of sand on the lake bottom." They are numerous along the west shore of Lake Huron. Some of these reach out into the lake a long distance and maintain their positions for years. They are also found along the shores of some of the inland lakes, such as Douglas Lake in northern Michigan and Silver Lake near Little Point Sable. That this type of cusp, with its underwater ridges, is built of material carried out from shore is shown by the changes that have occurred in Silver Lake. The level of Silver Lake changes slightly from year to year because of changes in its outlet. In those years when the water is low, it has a narrow sandy beach. Then cusps of Class 2 are always present. During the past summer, however, the water covered the sand, the vegetation along the shore prevented all beach drifting, and no cusps were formed.

Class 4[4] are "very small cusps formed at the lower ends of grooves which are sometimes formed just above the waterline and are perpendicular to it." These are always very small and never develop into cusps of a larger size. They are really a series of rill marks and, considering

[1] *Jour. Geol.*, Vol. XLVI (1938), pp. 615–27.

[2] *Ibid.*

[3] *Ibid.*

[4] *Ibid.*

their origin, probably should not be classed with beach cusps. However, since at the bottom of the channels are small cusplike deposits and since they have also been described as cusps by D. W. Johnson,[5] they should probably be retained in the classification.

Regarding Class 5 cusps,[6] which always occur in a series, long study and observation have convinced me that they never form except through the process of breaking of a ridgelike obstruction by the waves. This ridge is usually a beach ridge, but it is sometimes merely the edge of a low-cut bank. It is not necessary for a reverse slope to be present. All that is necessary is that the waves have an opportunity to cut through a long obstruction parallel to the shoreline. In the examination of hundreds of such series of cusps, I have found no exception to this. Also I have never seen a cusp series formed where the beach sloping toward the water was backed by an elevation too high for the waves to go over it.

The following example is typical of many beaches examined. On July 2, 1943, at Gray's Landing on Lake Michigan, about three miles north of White Lake Harbor, there were three series of cusps, each about 200 feet long, separated by equal distances of smooth beach. The waves had cut into the foredunes and caused the formation of a sand cliff several feet high. Where the cliff was close to the shore, the waves had built a smooth sloping surface which graded gently into the face of the cliff. Here no cusps formed, but in every case where the cliff retreated from the shore some distance, beach ridges, broken by cusp series, were present.

[5] *Shore Processes and Shoreline Development* (New York: John Wiley & Sons, Inc., 1919), pp. 475–76.

[6] Pp. 615–27 of ftn. 1 (1938).

RECENT STUDIES OF HALIBURTON-BANCROFT ALKALINE ROCKS: A DISCUSSION

FELIX CHAYES
Hyattsville, Maryland

This note is inspired by the recent publication of a geological map, by W. K. Gummer and S. V. Burr, of the area immediately east of Bancroft, Ontario.[1] The area shown on this map is the central part of a somewhat larger area that I had previously mapped and described.[2] The Gummer-Burr map is characterized by considerably less extrapolation than mine; and its location of specific outcrops or quarries is in general to be preferred, though there is, for the most part, little difference between the two maps.

Where the origin of rocks is so questionable, it seems very poor practice to introduce genetic nomenclature. Although the shonkinites and many of the syenites are probably "paragneisses," the use of the term as a map unit gives no indication of their peculiar mineralogical relation to the granites and nepheline rocks. (Most of the rocks I described as "potash-soda syenite-shonkinites" were collected from outcrops mapped as "paragneiss" by Gummer and Burr.) But this is a very minor matter; and all parties would be placated, I think, by the use of a composite term such as "syenitic-paragneiss" or "saturated-paragneiss." Despite the lack of significant difference between our maps, we are in almost complete disagreement concerning the origin of the alkaline rocks.

As far as field observation is concerned, the only significant difference between the earlier work of M. L. Keith,[3] my own study,[2] and the subsequent ones of Gummer and Burr,[4] J. Satterly,[5] and J. Thomson,[6] seems to me one of emphasis rather than of fact, hinging on the existence and importance of oversaturated or saturated rocks younger than the strongly undersaturated types. Although the common microperthitic syenites and shonkinites seemed to me to predate the nepheline rocks, I stated clearly that the plagioclase syenites were probably contemporaneous with them. The syenitic streaks which often cut across the foliation of gneissic foyaites I considered to have been derived by the reaction of liquid granite-pegmatite on solid nepheline rocks—a reaction for which the field offers considerable evidence. If these syenitic streaks are not of syntectic origin, then a certain amount of syenitic liquor must have remained liquid after the solidification of the nepheline rocks. Except for these rather insignificant streaks, I found no good evidence for such syenite; but Satterly has provided descriptions and an excellent photograph of outcrops in which gray syenite (the "plagioclase syenite" of my paper?) appears to carry inclu-

[1] *Ont. Dept. Mines Ann. Rept.*, Vol. LII (1943), accompanying report of J. Thomson.

[2] *Bull. Geol. Soc. Amer.*, Vol. LIII (1942), p. 452.

[3] *Bull. Geol. Soc. Amer.*, Vol. L (1939), p. 1795.

[4] *Science*, Vol. XCVII, No. 2517 (1943), p. 286.

[5] *Ont. Dept. Mines Ann. Rept.*, Vol. LII (1943), Part II, p. 13.

[6] *Ont. Dept. Mines Ann. Rept.*, Vol. LII (1943), Part III, p. 9.

sions of gneissic foyaite. Although there are numerous local developments of acid pegmatites younger than the nepheline rock, the results of my study were equally disappointing with regard to the existence of any considerable amount of younger granite. The intrusion of vast quantities of younger (but not necessarily related) granite is highly probable, but the few outcrops of massive granite I found gave no clear evidence of age relations. Since the massive granite is uncrushed, it could have appeared during or after deformation and hence could have accompanied or succeeded the nepheline rocks; the latter is surely more probable than the former.

But nearly all the granite of the area is gneissic, and most of both granite and the common microperthitic syenite shows evidence of mild deformation. In places the syenite has been thoroughly milled, and in the shonkinites evidence of deformation of the dark minerals is sometimes preserved despite their usually thorough recrystallization. Both gneissic granite and massive syenite are common as inclusions in the flow marble. In contrast, nepheline rocks have not yet been found as inclusions in the marble: nepheline in the nepheline pegmatites and injection gneisses shows evidence of neither "wet" nor "dry" metamorphism, and, concerning the massive foyaite of Blue Mountain, Keith[7] concludes that "there are no cataclastic structures or other microscopic evidence of movement or deformation during crystallization." Save for some of the acid pegmatites, no one has yet shown conclusively that any of the igneous rocks of the area are younger than the nepheline pegmatites which are the dominant undersaturated rocks of the eastern portion of the field; and it is my suspicion that such a demonstration is impossible. The issue between me and other recent students of the area thus resolves itself into (*a*) a question of fact, as to whether or not significant quantities of syenite and granite can be shown to be younger than the moderately undersaturated rocks; and (*b*) a matter of opinion, as to what importance should be attached to such a relationship. With regard to the question of fact small amounts of syenite seem to be younger than foyaite; and large quantities of granite *might* be younger, but my own evidence permits this inference only for scattered, isolated outcrops. What, then, of the second point?

The complexity of the Haliburton-Bancroft field is well known. From north to south there is a well-established transition from an area of igneous activity through a broad contact zone into a region of comparatively unaltered sediments. From west to east there seems to be a transition from truly plutonic igneous activity, as exemplified by Blue Mountain, to very late pegmatitic action, as shown so remarkably along the York River east of Bancroft. The plutonic nepheline rocks are foyaitic in composition; the pegmatitic ones, much more nearly urtitic. Except at Blue Mountain, very little of the magma crystallized *in situ*.

In a region as large and complex as this, it is certainly folly to base fundamental conclusions or objections either on the assumption that the magma composition varied uniformly in time throughout the field or on the parallel assumption that liquids of identical composition would crystallize simultaneously in different portions of the field. Both of these conclusions are necessary if the limestone-syntexis hypothesis is to

[7] P. 1824 of ftn. 3.

be rejected chiefly because occasional saturated or oversaturated rocks are shown or inferred to be younger than moderately undersaturated rocks. This seems the major premise of the argument of Keith, Thomson, and Satterly. It is to be expected that rock sequences will be contradictory with regard to intermediate members of the series, and it is surprising that the end members do not show similar relations.

A little syenite or granite later than massive or gneissic foyaites will not disprove the limestone-syntexis hypothesis any more than a little early syenite will prove it. But evidence suggesting that, in general, the oversaturated and saturated rocks are older than the strongly undersaturated rocks greatly increases the plausibility of the hypothesis; and evidence that some of the syenite, as well as scattered outcrops of massive granite, are or may be younger than the moderately undersaturated rocks will not seriously disturb this conclusion.

So much, then, for the differences of fact or emphasis between my work and that of Gummer, Burr, Thomson, and Satterly. Since my own work and that of Gummer and Burr were done in precisely the same area and we seem to conflict in matters of fact only to the extent already discussed, it will not be amiss to state my position concerning their theoretical conclusions.

That the nepheline pegmatites "may have . . . . been released by reactions of granitic or syenitic juices on limestone" seems plausible to me, but that "many of the nepheline pegmatites appear to be the result of regeneration of the nepheline of the paragneisses" seems to me without foundation. Rather, I suspect that its foundation is probably a study of many of the same outcrops from which I concluded the pegmatites were the source of the nepheline now found in the gneisses.

My own conception of the emplacement of nepheline[8] was simple and perhaps unimaginative: into the previously solid paragneisses a thin pegmatitic fluid of about urtitic composition was introduced. Where open cracks were available, they were filled (and perhaps greatly widened) by the liquid, which crystallized as nepheline pegmatite. Where the country rock did not fail by large-scale rupture, it was subjected to an extensive *lit-par-lit* injection of pegmatite liquids, the final product being a series of urtite-paragneiss hybrids.

Although this process might have been called "nephelinization" or "granitization," I felt that these terms could neither add to nor substitute for a more complete description and were therefore without any particular utility. The regeneration favored by Gummer and Burr is almost the exact converse of my explanation; yet the words "nephelinization" and "granitization" are as justly used by them as they might have been by me. What do we accomplish by such nomenclature?

Where contact planes are parallel to the structure of the wall rocks, most outcrops containing nepheline pegmatite and syenite, shonkinite, or foyaite are probably not incompatible with either hypothesis. There are localities in the eastern portion of the field, however, in which the interpretation of nepheline as anything but a late pegmatitic injection would scarcely be admissible. One of the best of these is the Snow Road Valley in the northeastern part of the town of Bancroft. The "regenerationist" may, of course, reply that nepheline injected into

[8] P. 503 of ftn. 2.

paragneiss at one locality has been derived *from* paragneiss elsewhere, but this sort of argument can be continued indefinitely.

It is one of the very great dangers of most "ultra-metamorphic" hypotheses that they cannot be definitely disproved. In this case, for instance, the mere fact that there is no presently recorded evidence for regeneration may be taken by some as an indication that regeneration has been so thorough as to destroy relations characteristic of initial and intermediate stages of the process. The only possible resolution of the controversy—one which is almost dictated by the over-broad terms in which it is necessary to state most "ultra-metamorphic" concepts—rests in the discovery of outcrops in which the distribution of nepheline can be explained only by its regeneration from paragneiss. Perhaps the brevity of the Gummer-Burr paper prevented inclusion of critical positive evidence of this type; it is to be hoped that the omission will be repaired. In any case, it is idle to speculate about the relative importance of regeneration as opposed to the more conventional *lit-par-lit* injection of liquid magmatic material before the presentation of at least a little field evidence in favor of the new suggestion.

PUBLICATIONS RECEIVED

Annotated List of Publications of the Department of Mines of the Province of Quebec, 1883–1944. Quebec: R. Paradis, 1944.

Annual Reports for the Years 1942 and 1943. State of Colorado, Bureau of Mines, 1943.

Archaeological Investigations in El Salvador. By John M. Longyear, III, with an Appendix by Stanley H. Boggs. Memoirs of the Peabody Museum of Archaeology and Ethnology, Harvard University, Vol. IX, No. 2. Cambridge, Mass., 1944.

Cheilotrypid Bryozoans from Pennsylvanian and Permian Rocks of the Midcontinent Region. By Raymond C. Moore and Ruth Mary Dudley. University of Kansas Publications, State Geological Survey of Kansas Bulletin 52, Part VI. Lawrence, 1944.

Common Rocks and Minerals of Missouri. By W. D. Keller. Arts and Science Series, No. 1. University of Missouri Bulletin, Vol. XLVI, No. 5. Columbia, 1945.

Connecticut State Geological and Natural History Survey. By Edward L. Troxell. Hartford: State of Connecticut, 1944. State Geological and Natural History Survey Bulletin 67.

The Correlation of Rocks of Simpson Age in North-Central Kansas with the St. Peter Sandstone and Associated Rocks in North Western Missouri. By Constance Leatherock. University of Kansas Publications. State Geological Survey of Kansas Bulletin 60, Part I. Lawrence, 1945.

Sur un crâne de crocodile fossile découvert dans le Miocene de Lisbonne. By A. Vianna et A. Moraes. Boletim da Sociedade Geologica de Portugal, Vol. IX, Fasc. III.

Dedication Proceedings of the Mineral Industries Building, Morgantown, West Virginia. West Virginia University Bulletin, Series 45, Nos. 3–11 (September, 1944). Morgantown University, 1944.

Department of Mines, Mining, and Geology, Biennial Report, 1943–44. Georgia State Division of Conservation, Geological Survey. Atlanta, 1945.

Diatomites of the Pacific Northwest as Filter-Aids. By Kenneth G. Skinner *et al.* U.S. Department of the Interior, Bureau of Mines Bulletin 460. Washington: Government Printing Office, 1944.

Dolomite Resources of Washington, Part I, Supplement. By Harold E. Culver, supervisor, Division of Geology. Report of Investigations No. 13. Olympia: State of Washington, Department of Conservation and Development, 1945.

Enough and To Spare. By Kirtley F. Mather. New York: Harper & Bros., 1944.

Etude des sondages de Kipula (Sud-Katanga). By A. Jamotte. Nouvelles observations sur le conglomérat de Mwashya et le petit conglomérat du Kundelungu. By P. Vanden Brande. Elisabethville, 1944.

The Excavation of Los Muertos and Neighboring Ruins in the Salt River Valley, Southern Arizona, Based on the Work of the Hemenway Southwestern Archaeological Expedition of 1887–1888. By Emil W. Haury. Papers of the Peabody Museum of American Archaeology and Ethnology, Harvard University, Vol. XXIV, No. 1. Cambridge, Mass., 1945.

Fifty Common Rocks and Minerals of West Virginia. By James H. C. Martens, mineralogist. West Virginia Geological Survey Bulletin 9. Morgantown, 1945.

Fourth Biennial Report of the State Department of Geology and Mineral Industries of the State of Oregon, July 1, 1942, to July 1, 1944. State of Oregon, Department of Geology and Mineral Industries Bulletin 28. Portland, 1944.

Geologic Literature of New Mexico through 1944. By Robert L. Bates and Marian R. Burks. New Mexico School of Mines Bulletin 22. Socorro, N.M.: State Bureau of Mines and Mineral Resources, 1945.

The Geology and Paleontology of the Marine Pliocene of San Diego, California, Part I: Geology. By Leo George Hertlein and U. S. Grant, IV. Memoirs of the San Diego Society of Natural History, Vol. II. San Diego, 1944.

Geology of the Cretaceous (Gualala Group) and Tertiary Formations along the Pacific Coast between Point Arena and Fort Ross, California. By Charles E. Weaver. University of Washington Publications in Geology, Vol. VI, No. 1. Seattle, 1944.

Geology of Water in Ohio: A Basic Report. By Wilber Stout, Karl Ver Steeg, and G. F. Lamb. Geological Survey of Ohio, 4th series, Bulletin 44. Columbus, 1943.

Ground-Water Conditions along the Ohio Valley at Parkersburg, West Va. By Russell M. Jeffords. West Virginia Geological Survey Bulletin 19. Morgantown, 1945.

Illinois Clays and Shales as Mortar Mix. By R. K. Hursh, J. R. Lamar, and R. E. Grim. Report of Investigations No. 100. Urbana: Illinois Geological Survey, 1944.

Illinois Surface Clays as Bonding Clays for Molding Sands. By R. M. Grogan and J. E. Lamar. Report of Investigations No. 104. Urbana: Illinois State Geological Survey, 1945.

An Introduction to the Archaeology of Cuzco. By John H. Rowe. Papers of the Peabody Museum of American Archaeology and Ethnology, Harvard University, Vol. XXVII, No. 2. Cambridge, Mass., 1944.

The Migration and Accumulation of Petroleum and Natural Gas. By F. M. Van Tuyl, Ben H. Parker, and W. W. Skeeters. Quarterly of the Colorado School of Mines, Vol. XL, No. 1. Golden, Colo., 1945.

Mining Review for the Half-Year Ended 31st December, 1940, No. 73. Mining Review for the Half-Year Ended 30th June, 1941, No. 74. Mining Review for the Half-Year Ended 30th June, 1942, No. 76. Mining Review for the Half-Year Ended 31st December, 1942, No. 77. Mining Review for the Half-Year Ended 30th June, 1943, No. 78. Mining Review for the Half-Year Ended 31st December, 1943, No. 79. Adelaide, South Australia: Department of Mines, 1941, 1943, 1944.

North Dakota Twenty-third Biennial Report. By Wilson M. Laird. Grand Forks, N.D.: North Dakota Geological Survey, 1944.

Optimum Producing Rates for Arbuckle Limestone Wells. By P. T. Amstutz, Jr., and Eugene A. Stephenson. University of Kansas Publications, Research Foundation Bulletin 1. September, 1944.

Pennsylvanian Morrowan Rocks and Fusulinids of Kansas. By M. L. Thompson. University of Kansas Publications, Bulletin 52, Part VII. Lawrence: State Geological Survey of Kansas, 1944.

Pennsylvania's Mineral Heritage: The Commonwealth of the Economic Crossroads of Her Industrial Development. Pennsylvania Department of Internal Affairs, Bureau of Statistics, Topographic and Geologic Survey. Harrisburg: Pennsylvania State College, School of Mineral Industries, 1944.

Permeability, Porosity, Oil, and Water Content of Natural Gas Reservoirs: Kanawha-Jackson and Campbells Creek Oriskany Fields. By A. J. W. Headlee and J. S. Joseph. West Virginia Geological Survey Bulletin 8. Morgantown, 1945.

The Physical and the Human Geography of the British Isles with a Map in 8 Colours. By Zofia Holub-Pacewicz. Glasgow, 1944.

Potash Salts from Texas–New Mexico Polyhalite Deposits. By John E. Conley and Everett P. Partridge. U.S. Department of the Interior, Bureau of Mines Bulletin 459. Washington: Government Printing Office, 1944.

Sur un principe de mobilité de l'émanation ou de l'intrusion granitiques. By A. Jamotte. Note sur la probabilité de l'existence d'algues fossiles du genre *Collenia* dans la série des mines du Katanga. By A. Jamotte. Nouvelles observations sur la presénce d'organismes du genre *Collenia* dans la série des mines. By A. Jamotte and P. Vanden Brande. Sur la découverte de *Collenia* sp. dans le system schisto-dolomitique et sur le problème de la série des mines du Katanga. By A. Jamotte. Notes complémentaires sur la stratigraphie de la série des mines et sur *Collenia* sp. By A. Jamotte. Elisabethville, 1944.

Rates of Sediment Production in Southwestern United States. By Carl B. Brown. U.S. Department of Agriculture, Soil Conservation Service, January, 1945. Mimeographed.

Report of the Geological Survey for the Years 1941–42. Perth: Robert H. Miller, Government Printer, Western Australia, Department of Mines, 1944.

Report of the State Geologist on the Mineral Industries and Geology of Vermont, 1943–44. By Elbridge C. Jacobs, state geologist. Burlington, 1944.

Report on an Investigation of Water Losses in Streams Flowing East Out of the Black Hills, South Dakota. By Carl B. Brown. Special Report No. 8. Washington, DC.: U.S. Department of Agriculture, Soil Conservation Service, October, 1944.

A Résumé of Paleozoic Stratigraphy of Colorado. By J. Harlan Johnson. Quarterly of the Colorado School of Mines, Vol. XL, No. 3. Golden, Colo., July, 1945.

Review of Petroleum Geology in 1944. By F. M. Van Tuyl and Others. Quarterly of the Colo-

rado School of Mines, Vol XL, No. 2. Golden, Colo., 1945.

A Revised Glossary of Queensland Stratigraphy. By W. H. Bryan and O. A. Jones. University of Queensland Papers, Department of Geology, Vol. II (new ser.), No. 11. Brisbane, 1944.

Selecnium: In Glacial and Associated Deposits. By Walter V. Searight and Alvin L. Moxon. Agricultural Experiment Station, South Dakota State College of Agricultural and Mechanic Arts. Brookings, S.D., March, 1945.

Selected Well Logs in the Virginia Coastal Plain North of James River. By D. J. Cederstrom. Virginia Geological Survey Circular 3. Prepared in co-operation with the Geological Survey of the U.S. Department of the Interior. University, Va., 1945.

Sillimanite and Massive Kyanite in Georgia. By A. S. Furcron and Kefton H. Teague. Georgia State Division of Conservation, Department of Mines, Mining, and Geology, Bulletin 51. Atlanta, 1945.

Some Magnetite Deposits of Stevens and Okanogan Counties, Washington. By W. A. Broughton. Report of Investigations No. 14. Olympia: Department of Conservation and Development, Division of Geology, 1945.

Stratigraphic and Paleontologic Studies of Wells in Florida, No. 4. By W. Storrs Cole. State of Florida, Department of Conservation, Florida Geological Survey, 1945.

The Stratigraphy of Some Lower Ordovician Formations of the Ozark Uplift. By James S. Cullison. University of Missouri School of Mines and Metallurgy Bulletin Technical Series, Vol. XV, No. 2. Rolla, 1944.

Stratigraphy and Paleontology of the Tertiary Formations at Coos Bay, Oregon. By Charles E. Weaver. University of Washington Publications in Geology, Vol. VI, No. 2. Olympia, June, 1945.

Stratigraphy and Structure of North Dakota. By Wilson M. Laird. North Dakota Geological Survey Bulletin 18. Grand Forks, 1944.

A Survey of Weathering Processes and Products. By Parry Reiche. University of New Mexico Publications in Geology No. 1. Albuquerque, 1945.

Tulsa Geological Society Digest. By Roy L. Ginter, editor. With abstracts of papers presented before the society from October, 1943, to June, 1944. Tulsa Geological Society, 1945.

The Underground Water of the Southeastern Part of South Australia. By L. Keith Ward. Department of Mines Geological Survey of South Australia Bulletin 19. Adelaide, 1941.

MAPS RECEIVED

Mineral Resources of Missouri. Missouri Geological Survey and Water Resources. Scale: 10 miles to the inch. Rolla, 1944.

Preliminary Map, Prosperous Lake, Northwest Territories. By A. W. Jollife. Canada Department of Mines and Resources, Mines and Geology Branch, Geological Survey Paper 45-4. Scale: 2 inches to 1 mile. Ottawa, 1945.

Physiographic Diagram of Asia. By A. K. Lobeck. Scale: 1:20,000,000. New York: Geographical Press, Columbia University, 1944.

First Generalized Geologic Map of South America. By Victor Oppenheim. United States Section Pan American Institute of Mining Engineering and Geology. Technical Paper No. 2. Scale: 1:11,200,000. New York, 1945.

REVIEWS

Foundations of Plant Geography. By STANLEY A. CAIN. New York: Harper & Bros., 1944. Pp. xiv+556; figs. 63. $5.00.

The scope of a scientific field increases so rapidly that it is next to impossible for an individual worker to maintain a proper perspective not only of his own field but also of all related fields. Therefore, we are greatly indebted to the author for the completion of a herculean task: the assembly of a large amount of material, in what he chooses to call *The Foundations of Plant Geography*, and its analysis and evaluation. Analyses of groups of related titles are in many cases deeply penetrating. The conservative nature of the evaluation accomplishes two things: it leaves the reader with a desire for more extended evaluation, difficult as the task may be, and it creates an urgent stimulus to further investigation by the clean-cut delineation of a myriad of problems.

In addition to the text, Dr. Cain's book contains a glossary of uncommon terms, a literature-cited section of 720 titles, and a rather complete index. The text contains thirty chapters divided into five parts as follows: "Introduction," "Paleoecology," "Areography," "Evolution and Plant Geography," and "Significance of Polyploidy in Plant Geography." By no means does the book conform to the ordinary idea of descriptive plant geography. If one desires a simple answer to the problem, "Why plants are where they are and how they got there," he will be disappointed. The book, rather, is a critical review of a large mass of literature excellently organized. Although the temptation exists to question the absence of certain titles from the Bibliography, a single book is necessarily restricted to the objectives of the author, and it is obviously impossible to include all work considered pertinent to plant geography by all workers. In the opinion of the reviewer, Dr. Cain has performed a highly valuable service, one which might well be emulated in other fields from time to time.

The text material dwells heavily upon paleoecology, dispersal, speciation, and polyploidy. Throughout its entirety the central theme appears to be evolution, and this viewpoint binds together contributions from paleobotany, taxonomy, genetics, cytology, dynamic plant distribution, and the causes of speciation. As a matter of fact, an additional background of plant physiology, plant anatomy, paleobotany, physiography, climatology, meteorology, and pedology is a great help for the full appreciation of Dr. Cain's review. The whole work illustrates clearly how progress in a single field of investigation depends upon and is assisted by contributions from many fields.

Some readers might take exception to the title of "Foundations" or even of "Principles." However, the principles of plant geography constitute the entire theme of chapter ii and are emphasized from point to point in Part II, "Paleoecology." The principles are well summarized with good argument and pertinent examples except, perhaps, in the case of optimum and limiting factors, which are rather involved topics for the restricted consideration given them.

Because of the manner of treatment, geologists and paleobotanists are most likely to be interested in the large section devoted to paleoecology, including chapters on "The History of Cenozoic Vegetation of Western America" and "Pollen Analysis as a Paleoecological Research Method." Emphasis is placed upon principles, methods, and examples from recent work, especially from the penetrative and comprehensive work of Chaney. In spite of a slight perplexity in the author's mind as to the balance between divergent interpretations, his critical analyses are straightforward and lucid. One wishes he had been able to go even further in the selection of salient features of various schools of thought and their evaluation. The extensive work of Chaney and his students, of Mason, Berry, and others, merits an analysis and evaluation beyond the space permissible in the present book. In any event, if a person wishes to be sufficiently well grounded in the principles of paleoecology and their application, he must necessarily digest the details in the original articles. The brevity of the one chapter on pollen analysis, although a myriad titles are concisely interwoven, is no doubt justified in the light of the excellent books already in existence.

The remainder of the book tends increasingly toward an evolutionary approach. At times the connection with plant geography becomes tenuous or even appears to be lost sight of altogether. This is especially true of such topics as the nature of species, speciation, species stability, and polyploidy. The diverse topics could, perhaps, have been more intimately interwoven—a most difficult task. However, the reader perceives a future value of high order to plant geography in research suggested by Dr. Cain's treatment. A wealth of well-co-ordinated information, especially in the topics of area, dispersal, centers of origin, endemism, senescence, and isolation, yields a broad perspective not only for those who work in related fields but also for those directly concerned with the topics. This, of course, is quite true for the entire book.

On the whole, we are indebted to Dr. Cain for accomplishing three things: first, he has assembled, reviewed, and evaluated a large amount of the recent literature pertinent to plant geography or related in some way thereto; second, he has focused attention once again upon the essential unity among several branches of science; and, third, he has in numerous cases highlighted the many problems awaiting investigation.

WALDO S. GLOCK

Foreign Maps. By EVERETT C. OLSON and AGNES WHITMARSH. New York: Harper & Bros., 1944. Pp. xvii+237; figs. 25; tables 7; pls. 16. $4.00.

The present conflict is a war of maps as well as of ships, tanks, and planes. Soldier and civilian alike have become involved, it seems at times almost submerged, in the flood of maps flowing, no longer by hundreds and thousands, but by tons and carloads. Numerous manuals and other books have appeared as aids to instruction in the making and reading of a great variety of maps. *Foreign Maps* is the latest of these and is almost the only volume in English which, for some forty countries, presents a brief but comprehensive treatment of their medium- and large-scale topographic maps. It is to be hoped that in the second edition large maps on a scale approximating an inch to a mile, which together might be considered a master-map of the world, will be treated more fully.

The primary organization of the book is topical, with chapters on signs and symbols. scales and measurement, marginal information, map indexes, and grid systems. Its unique features are proposals on procedure in reading foreign maps, an analysis of the characteristics of maps issued by national mapping agencies, and an extensive treatment of the difficult problem of language.

More than eighty pages are devoted to the translation into English of map vocabulary from Chinese, Japanese, German, Russian, Turkish, even Tibetan, and thirty other languages. The principal barrier to the use of foreign maps has been breached successfully.

Sixteen samples of foreign maps, half of them in color, are included in the volume. They are well chosen to illustrate various types of treatment, though clarity has in some cases suffered in reproduction. An arrangement whereby more or all of the maps could be seen at one time, addition of marginal data, and retention of original scales would improve their usefulness.

Foreign Maps is a timely and original description and analysis of the medium- and large-scale series of maps issued by most of the national mapping agencies of the world.

H. M. LEPPARD

The Structural Control of Ore Deposition in Some South Australian Copper Fields. By S. B. DICKINSON. Department of Mines, Geological Survey of South Australia, Bulls. 20 (1942) and 21 (1944). Pp. 99; figs. 29; and pp. 66; figs. 14.

Part I of Bulletin 20 deals with the Wallaroo-Moonta field, which, since its discovery in 1860, has been the most productive copper-mining field of South Australia, yielding copper ore roughly estimated at six and a quarter million tons. Large-scale mining ceased in 1923 and the mines are now completely dismantled. The only previous geologic report on the district is that of Dr. R. L. Jack, general in scope and published in 1917. The present report is restricted to the structural geology and is based wholly on the examination of mine maps and other office records, since direct underground examinations were impossible and surface studies were useless because a blanket of post-mineral formations completely masks the ore-bearing horizons. It is hoped that the present structural studies will offer aid and encouragement to further exploration for ores.

The ore deposits were formed in pre-Cambrian times and are hypothermal vein deposits traversing pre-Cambrian acid igneous rocks and metamorphosed sediments. In the Moonta district the ores occur in porphyry along two sets of thrust faults striking northeast and southwest and dipping to the northwest. The two sets diverge about 15° in strike. An unusual technique in some of the stope maps is the representation of the workable widths of certain ore bodies by contour lines. This brings out very well the pipelike forms of some ore bodies and the tabular form of others.

Cross-faults intersecting the principal veins are some of them pre-mineral and some post-mineral. The former have influenced mineralization by blocking mineralizing solutions, and the latter have displaced pre-existing ore bodies.

In the Wallaroo district 9 miles to the north, the ores occur in pre-Cambrian sedimentary and metamorphic rocks of whose rock structure nothing is known. Unlike the Moonta district, the ore bodies lie in steeply dipping east-west faults and have been sliced and mildly displaced by northeast-southwest shear zones. The latter trend similarly to the principal veins of the Moonta district, with which they may be contemporary, but they do not contain workable ore bodies. The author relates the fracture pattern in both the Moonta and the Wallaroo districts to various elements of the same strain ellipsoid.

Part II on the Dome Rock Copper Mine is a complete geologic study of an active copper mine characterized by small but interesting ore deposits. The mine lies 26 miles north of Mingary siding. The area consists of metamorphosed sandstones, shales, and ferruginous sandstones intruded by granite. Although the mine workings are within a few hundred feet of the border of the granite batholith, the minerals indicate mesothermal conditions of deposition. The sediments are lithologically similar to those at Broken Hill and have usually been regarded as pre-Cambrian. The age of the granite is unknown.

The two principal ore deposits are of especial interest because they are small pencil-like, steep-pitching patches of ore localized in minor drag-fold structures on the overturned limb of a strong anticline. To a depth of 200 feet the ores are oxidized. The primary ore consists chiefly of pyrite and chalcopyrite, which replace the country rock, usually a fine-grained sandstone.

Part III describes copper deposits of the "Red Beds" type found in the Mountain Gunson–Pernatty lagoon district. These deposits are localized within a distinctive flat-lying bed of dolomitic limestone and in conformable red sandstones immediately above or below. The cupriferous sediments are nonfossiliferous and appear to be terrestrial. They have been variously classed as Ordovician and as pre-Cambrian. The principal commercial deposits are at the Mount-Gunson mine, from which the ores are shipped to the Port Pirie smelters. Their high silica content gives them fluxing value in the smelting of less siliceous ores. In a few localities sulphides are present, chalcocite, covellite, and bornite; but most of the ores are oxidized. The irregular distribution of the mineralization within the sediments indicates an epigenetic origin. Conclusive evidence is lacking as to whether the mineralizing solutions were meteoric or magmatic (hydrothermal) in source. The reviewer cannot agree with the statement that the presence of barite and fluorite in some of the ores indicates that the ore solutions were hot rather than cold. There is ample evidence that both minerals may be deposited from cool meteoric waters as well as from thermal waters.

Part IV deals with the Burra Burra mine about 100 miles north of Adelaide. The mine was discovered about 1851 and worked for 26 years, yielding copper to the value of nearly five million pounds sterling. The rocks of the area consist of a conformable series of folded and mildly metamorphosed sediments that are without fossils and usually classed as pre-Cambrian. A single feldspar-porphyry dike intrudes these sediments.

The Burra Burra ore body was about 800 feet long and 250 feet in maximum width. Below a superficial leached zone the ore body extended to depths of only 300 feet. It was, however, very rich, running from 20 to 30 per cent copper for the dressed ore. All the commercial ore consisted of oxidized copper minerals and some chalcocite, but exploration by drilling has shown the existence below of pyrite and chalcopyrite in scattered distribution and not of commercial grade.

The ore resulted from processes of oxidation and downward enrichment acting on a pyrite-chalcopyrite protore above and near the groundwater level in an area of dolomites and calcareous shales shattered by faulting.

Parts V–IX relate to copper-producing dis-

tricts of lesser importance. They are of historic and local interest mainly.

EDSON S. BASTIN

Asia's Lands and Peoples: A Geography of One-third the Earth and Two-thirds Its People. By GEORGE B. CRESSEY. New York: McGraw-Hill Book Co., Inc., 1944. Pp. xi+608. $4.50.

This well-timed text is the first on Asiatic geography to present the continent in something like the perspective it should have in the middle of the twentieth century. Dr. Cressey has been in a peculiarly favorable position to bring the picture into better balance than has been done before, first by protracted residence in China and then by his unique good fortune in serving as adviser to the government of the U.S.S.R. during the preparation of the *Great Soviet World Atlas* (1938–40), material for which was inspected by delegates to the International Geological Congress at Moscow in July, 1937. In the ensuing months he had access to a wealth of new data and a variety of experiences which any geographer might envy. The result is the 120 pages in the heart of the volume which form its outstanding contribution.

The book is replete with factual material, much of which has not yet found its way into standard texts. Cressey's section on the natural resources of the U.S.S.R. is based on the figures for 1940. His Japanese estimates are as up to date, if not necessarily equally reliable.

The plan of the book is sound and well balanced. It abandons the classical line of attack from the western flank and approaches from the southeast, after a preliminary briefing in Hawaii on the geostrategy of the Pacific Basin. The division of the continent into major realms essentially follows Bergsmark's of 1935. Descriptions of each region start with a general picture of the area, followed by a sketch of the physiographic, climatic, and pedological foundations, upon which thereafter the distribution of vegetation, agriculture, and social life develop naturally. Good sentence-summaries of significant geological structures appear where germane to the discussion of land-forms and natural resources. Pertinent historical background is introduced in various places, though not always with full appreciation of the factors involved. Thus, while it is true that in early historic times the frontier of Japanese expansion stood for long on the shores of Lake Biwa, it should surely be added that just across the lake lay the real barrier, the tough rampart of the Central Mountain Knot.

The style is readable, often vivid, at times with arrestingly epigrammatic turns of phrase. Of the Soviet realm, "too much of the land is too cold, or too dry, or too wet . . . . or too something else. . . . . It has the longest coastline of any country, and the most useless. . . . Even the rivers flow in the wrong direction." But the reviewer deprecates such verbal plastic surgery as the excision of prepositions from the subtitle of the book; even quotation marks would hardly atone for this journalese liberty!

The book strikes a fresh note in another respect. No study of the adjustment of man to his environment can afford to leave out of account racial psychological factors, especially where these influence national aspirations toward expansion. This is peculiarly true of Japan, where failure to allow for this driving force behind the people of the island empire led to their being seriously underestimated as a potential foe. (In this connection the map showing the progressive extension of Japanese hegemony is instructive to others besides students of history.) The chapter on "Japan's World Position" brings this whole aspect into proper focus, even though it is surprising to find no reference to the counterbalancing psychological "tug of the homeland," which has prevented the race from making good colonists and has limited the energies of the overseas Japanese on the Asiatic mainland to exploiting the localities in which they sojourned in hope of an early return to the islands of their birth.

It is to be regretted that, with so much to commend it, the publication shows so many signs of hasty preparation. Even with the plan, stated in the Preface, of confining to certain chapters the high lights of the major political units and relegating factual details to other sections of the book intended for more advanced students, the impression is left that the author never quite decided whether he was writing a textbook or a reference book. There is an alternation between excellent descriptive passages and masses of almost ponderous detail which must be studied with an atlas to derive their real value.

Nor are different sections of the book always thoroughly co-ordinated. Little is gained by introducing two independent classifications of the physiographic provinces of the same region, and

the result can only be confusing to the student. Thus the map on pages 270–71 shows "Landforms of the Soviet Union," devoid of the key-letters which would link them instantly with the systematic listing on the previous page; while pages 312–13 present "Geographic Regions and Geomorphic Realms" of the same area, indexed by a different letter-system and with different boundaries, yet with no attempt to reconcile apparent discrepancies. Two different climatic classifications of Japan appear on the same page. Similarly, the physiographic regions of Tibet are grouped differently on pages 55 and 156, with no explicit correlation between the nine numbered regions and the seven lettered ones of the same area and with a misplaced mountain range flanking the Karakorum. In other cases glaring errors must be due to slipshod editing—as where, among "India's Varied Flora," the astonished reader finds "tidal forests" mapped along the crest of the Himalayas and "dry alpine vegetation" along the Karachi coast. The "cyprus" trees, "Twai" River of China, and the like are evidently compositor's novelties.

The illustrations are, for the most part, culled from fresh sources and include a number of good air-photos, though the halftones fail to do them justice. All the diagrams and charts were specially drafted, but here, too, inadequate supervision of the work mars the result. A binding hardly robust enough for the heavily weighted paper may be a wartime necessity; but the value of diagram maps need not have been sacrificed by the use of confusing or indistinct conventional shading, unbalanced weight of boundary lines and lettering, or the absence of reference points on distribution maps. The three-dimensional effect of Raisz's characteristically beautiful Harvard-Yenching Physiographic Diagrams is largely killed by the superposition of crudely drawn boundaries. This is partly due to overreduction of the originals, a defect seen also in the blank areal maps, large margins of which might have been cropped as an alternative to excessive condensation of the significant region. While in comparing different regions there is obvious advantage in a uniform scale, that purpose could be achieved by a key-map on 1/30,000,000 at the entry to each new region, after which the scale of distribution-maps could be determined by the nature of the detail to be exhibited. Thus the map of land-forms of the Japanese island chain could be reproduced on a much larger scale by "insetting" the extreme northern outposts and Formosa, instead of including blank expanses of all Manchuria, southeastern Siberia, and 500,000 square miles of China and Mongolia, so that a lens is needed to distinguish the two symbols which alone justify the inclusion of the map. No key is given for the most prominent symbol (*SB*) on the map of China's raw materials, and the student must wait until the next chapter before guessing that it means "soybean."

Mention of such defects may seem petty; but where the aim of a book is to throw a fresh image on the mental screen, details of projection become important. It is only because these details can be readily corrected in a second edition that it is worth pointing them out. In a book of less genuine merit they could pass without comment.

A useful addition is the extensive annotated Bibliography which lists with discrimination all the more important publications in English up to 1943, as well as a number of standard ones in other languages. A valuable inclusion is the reference to available maps of the various regions; and instructors will note with appreciation that lantern slides of the photographic illustrations are securable from the Syracuse University bookstore.

George B. Barbour

INDEX TO VOLUME LIII

INDIAN AGRICULTURAL RESEARCH INSTITUTE LIBRARY, NEW DELHI.

Date of Issue	Date of Issue	Date of Issue

GIPNLK—H-4 I.A.R.I.—29-4-55—15,000